ANTILLEAN VEGETATIONAL HISTORY AND PALEOCLIMATE
RECONSTRUCTED FROM THE PALEOLIMNOLOGICAL RECORD OF
LAKE MIRAGOANE, HAITI

By

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DEDICATION

I affectionately dedicate this work to my professor, the late Dr. Edward S. Deevey Jr., and to my dear and loving parents, Evaristo Higuera P. and Maria N. Diaz.

ACKNOWLEDGEMENTS

continuous advice and encouragement truly made the task of finishing this work a reality. Tim provided me with encouragement and emotional support and for the past three years, he has covered all the expenses related to this work. Thus, for the kindness and unsurpassed support I received from Mark and Tim, I wish to express to them my most sincere appreciation.

Thanks to the professional care I received from Dr. William Friedman, I can once again enjoy the pleasure of a painless life. His accurate treatment enabled me to return to what I was before hurting my back: a healthy, happy, hard- working person who likes to smile.

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Palynological analyses of sediments from Lake Miragoane, Haiti, provide a
continuous record of the vegetational and climatic history of Hispaniola from the
late Pleistocene to present. The demonstrated high sensitivity of Antillean plant
communities to long-term climatic fluctuations has profound implications for West
Indian biogeography. Available isotopic ($^{18}O/^{16}O$) and geochemical records from
Lake Miragoane were used in combination with the pollen results for reconstructing
the ecological history of the lake and its watershed.

Extremely dry conditions characterized Hispaniola during the late Pleistocene,
$\geq$ 10,230 - 9,700 B.P., and early Holocene, 9,700 - 7,000 B.P., with temperatures
cooler than today persisting until $\approx$ 8,600 B.P. During these dry phases, local and
regional forest fires were rare. The vegetation around Miragoane, and possibly
throughout Hispaniola, consisted of xeric and shrubby communities and small,
fragmented forests. Lake Miragoane, with permanent fresh water standing $\approx$ 10 m
below present as the Pleistocene ended, filled when the Holocene began. Deep-

water conditions and a large littoral zone occupied by Chenopodiaceae/Amaranthaceae persisted until 2,800 B.P.

Hispaniola's most mesic vegetation flourished in the mid Holocene (7,000 - 2,800 B.P.) under a wet, warm, seasonal climate. Successional mesic forests predominated until 4,800 B.P. and later co-existed with mature forests until 2,800 B.P. Forest fires became common in Hispaniola by 7,000 B.P.

Climatically dry conditions returned around 2,800 B.P. Since then, the watershed's vegetation has consisted of dry communities and relict moist forests. Agricultural activity at Miragoane dates from 1,100 B.P., but the greatest deforestation took place during the last five centuries.

Lake water levels, though fluctuating, have drastically declined since 3,700 B.P. Lacustrine sedimentation patterns in the past millennium have been strongly controlled by declining water levels, establishment of macrophyte beds, and human disturbance. A large marsh on the basin's east side has probably functioned as a sink for incoming clastics and as a net source of organic matter to the lake. Consequently, slow deposition of organic-rich sediments has coincided with severe human-induced disturbance of the watershed.

INTRODUCTION

Evidence of late Quaternary climatic fluctuations in the Caribbean region, like elsewhere in the tropics, is beginning to emerge, but the impact of these changes on the Antillean terrestrial ecosystems remains unknown. Information on long-term climate changes in the Caribbean derives from pollen sequences from mainland sites (Wymstra and Van der Hammen, 1966; Leyden, 1984, 1985; Bartlett and Barghoorn, 1973; Watts, 1980, 1971; Watts and Stuiver, 1980), from studies of isotopes and/or foraminifera of deep-sea cores from the Caribbean/Gulf of Mexico basin (Lynts and Judd, 1973; Keigwin and Jones, 1989), and from paleoclimate modelling of the region (Overpeck et al., 1989). Morphological changes observed in Jamaican fossil land snails provide evidence of late Pleistocene and early Holocene dry conditions (Goodfriend and Mitterer, 1988), as well as indications of a drying climatic trend during the past two millennia (Goodfriend, 1987). In general, these records suggest aridity and cold temperatures in the late Glacial and moisture fluctuations during the warmer Holocene in circum- Caribbean lowlands.

Evidence of glaciation in Hispaniola is documented by morphological and sedimentological features (e.g. cirques, moraines, bog-filled depressions, etc.) present in the Cordillera Central, Dominican Republic (Schubert and Medina, 1982). Early Holocene eustatic sea level rise has been linked to flooding and subsequent formation of coral reefs in the Enriquillo Valley, Dominican Republic (Mann et al., 1984; Taylor et al., 1985), and to the development of coastal wetlands and the offshore shelf in Jamaica (Hendry and Digerfeldt, 1989). A climatic record

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spanning the last 20,000 years was reconstructed from mineralogical analyses of cave guano deposits from Mona Island, Puerto Rico (see Carbone, 1978). The climatic sequence (which presumably documents the Little Ice Age event of northern latitudes) is, for the most part, in strong conflict with the changes in climate inferred from adjoining localities, including those reported from Hispaniola (this work).

The sediment core studied for this project was raised from Lake Miragoane, located in Haiti's southwestern peninsula (Fig. 1). Some preliminary pollen results of this core have been already published (Higuera-Gundy, 1988), as are several papers on human-induced disturbance of the drainage basin (Higuera-Gundy et al., 1986; Binford et al., 1987; Brenner and Binford, 1988; Higuera-Gundy 1989), and an $^{18}O/^{16}O$ record of lake level fluctuations (Hodell et al., 1989, 1991 in review).

Lake Miragoane was chosen as a sample site to study the paleoecology of the West Indies after performing exploratory visits and preliminary coring to determine its suitability for this kind of research. Following bathymetric and limnological studies, it became evident that this fresh-water lake, perhaps the largest in the Caribbean islands, could contain a well-preserved paleoenvironmental record, old enough to pursue our goals.

The primary objectives set forth at the initiation of this multidisciplinary project were 1) to investigate the effects of late Glacial and Holocene climatic fluctuations on the vegetation of Hispaniola and, by inference, on the remaining Antilles, and 2) to assess the impact of human-induced disturbance on Miragoane's local flora. For these purposes, palynological analyses were performed on sediment samples from a 7.67-m core taken from the central, deepest (41 m) part of the lake (Fig. 1). Radiocarbon dating of bulk organic matter in sediments at 7.53 m rendered an age of 10,230 ± 160 B.P.

The palynological data of Lake Miragoane demonstrate that the insular plant communities have been as susceptible to long-term climatic changes as their continental counterparts. From late Pleistocene to present, a sequence of dry-mesic-dry climatic/vegetational cycles is identified, which coincides with episodes of lake water level fluctuations (Hodell et al., 1989, 1991 in review). Some near-shore vegetational changes have indirectly contributed to alter the composition of lacustrine sediments and the patterns of sedimentation in the lake, according to geochemical data (Binford et al., 1987; Brenner and Binford, 1988). The charcoal record documents a complex history of changes in fire frequency related to natural (climatic and non-climatic) and anthropogenic factors. Agricultural activity within Miragoane's watershed is a recent event, dating back no more than 1,500 years, though the most severe disturbance of the local soils and flora has occurred in the past 500 years and continues, with even greater severity, at present.

Changes in climate/vegetation inferred from Lake Miragoane are compared with findings in other sites in an attempt to assess the Antillean archipelago's response to climatic oscillations, as these islands extend over a wide latitudinal range. Episodes of major vegetational change are discussed in relation to periods of faunal extinctions in the Antilles. Historical and archaeological data, as well as other relevant sources of information, were also incorporated when reconstructing Miragoane's (and Hispaniola's) environmental history.

DESCRIPTION OF THE SITE
Geographic Setting

Lake Miragone lies in limestone terrain about 5 km from the north shore of Haiti's southwestern peninsula (Fig. 1). The basin sets in a tectonic rift system at 18°24' N, 73°05' W. Lake Miragoane is hydrologically connected to a smaller lake, Petit Etang de Miragoane, through an eastern marsh ($\approx$ 2 - 3 km long). These two lakes, along with a smaller, unnamed pond and a marine embayment of the Caribbean sea constitute the Miragoane Lakes basin, which evolved as a result of pull-apart activity of the sinistral Enriquillo-Plantain Garden Fault Zone (EPGFZ) in Haiti (Mann et al., 1983).

Morphometry and Limnology of Lake Miragoane

The bathymetric map of Lake Miragoane (Fig. 1) shows a nearly circular basin, only interrupted on the southwest shore by a shallow (z_{max} = 25 m), conical bay. With a surface area (A) of 7.06 km^2 and mean water conductivity of 375 μS/cm, Miragoane may be the largest, natural fresh water lake in the Caribbean (Brenner and Binford, 1988). Because the lake's maximum depth (z_{max}) is 41 m and the water surface lies at 20 m above m.s.l., the basin has a 21 m cryptodepression. It is suspected that Lake Miragoane may be a warm monomictic system. Thermal stratification (well defined during the summer) may begin around April, while water

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Figure 1. Map of the Caribbean region (modified from Hodell et al, 1991 in review). Vegetational map of Hispaniola showing the approximate distribution of dry, moist, and pine forests (modified from Holdridge, 1945). Bathymetric map of Lake Miragoane, Haiti, showing 10-m contour depths (modified from Brenner and Binford, 1988) and cored site. Lake Miragoane lies 5 km from the north shore of Haiti's southwestern peninsula. The nearest mountains to the lake are Massif de La Selle, to the east, and Massif de La Hotte, to the west.

CARIBBEAN REGION
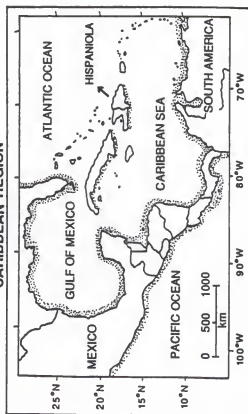
HISPANIOLA
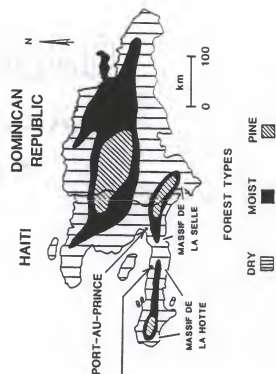
LAKE MIRAGOANE

FOREST TYPES

DRY   MOIST   PINE
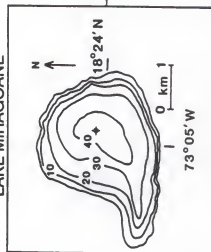
circulation may begin in late December or early January (Brenner and Binford, 1988).

Temperature and oxygen profiles measured on August 27, 1983, and again on July 28, 1985 are similar (see Brenner and Binford, 1988; Binford et al., 1986). Data taken in 1985 indicate that water temperatures declined from 28°C at the surface to 22°C at 40 m. The zone of maximum temperature drop (from 26.2 to 24.8°C) was detected between 20 and 25 m.

The summer oxygen curve was clinograde, with higher O_2 concentrations in the epilimnion than in the hypolimnion, and a sharp decrease in oxygen content along the thermocline. Oxygen concentrations in the surface 15 m of water were nearly constant, 7 mg/l. Between 15 and 25 m water depth, the O_2 content dropped from 7 mg/l to < 1 mg/l. Below 25 m, the water became quickly depleted of oxygen, reaching anoxia at the bottom (Brenner and Binford, 1988).

Climate

Haiti's climate is warm and equable. Mean annual temperatures in coastal lowlands vary between 25°C and 27°C, whereas the highlands are somewhat cooler. The mean annual temperature decreases at an estimated rate of 1°C per every 150 m increase in elevation, though snow and ice probably do not occur in the Republic (Woodring et al., 1924).

Annual rainfall is distinctly seasonal, with the bulk of precipitaion falling in late spring and early autumn. These two wet periods are separated by a long dry winter season, and a short midsummer episode of drought. The spring rains last from April through June, reaching a peak in May. July is the driest and, usually, the hottest summer month. The autumn rainy season is longer and begins at different

times in different places, though on the average, October and November have the greatest precipitation. In general, the highlands receive nearly 50% more rain than lowland sites (Woodring et al., 1924). The long dry winter season tends to be quite favorable for the occurrence and spread of forest fires (Holdridge, 1947), despite having mean temperatures $\geq 2°C$ cooler than the summer months.

In the summer, the northward displacement of the equatorial storm belt causes the precipitation to arrive as heavy, short-lasting thunderstorms. From October to March, the rains brought in by the northeasterly winds are much lighter, steady, and last from several hours to two or three days, but few deliver large amounts of water (Woodring et al., 1924).

Spatial distribution of rainfall in Hispaniola is rather irregular due to the island's rough topography. Much of the rain delivered to the island by the trade winds, blowing from the east and northeast, is intercepted by the high mountains of the Dominican Republic. The Republic of Haiti lies in rain shadow to the west of these mountains and is, therefore, considerably drier than the Dominican Republic (Woodring et al., 1924).

Peninsular sites in the vicinity of Lake Miragoane are, like the rest of Haiti, warmer and drier in the lowlands and cooler and moister in the uplands. Precipitation in coastal lowlands near Lake Miragoane averages 1,000 to 2,000 mm per year, and is somewhat greater along the northern shore than in the southern coast (Woodring et al., 1924). Mean annual rainfall in the mountains varies from 2,000 mm in the Massif de La Selle, to the east of the lake, to 3,000 - 5,000 mm in the Massif de La Hotte, to the west (Woods, 1987).

Vegetation

A broad classification of Hispaniola's rich flora has been sketched by Holdridge (1945). According to the author, plant associations typical of the island include lowland dry forest, highland mesic and pine forests, and coastal mangrove swamps (Fig. 1). Though these categories provide an insight into the island's main vegetation, further subdivisions of the insular plant communities are likely to result as floristic studies continue.

Dry vegetation is mostly confined to the warmer and drier coastal lowlands, and can consist of dry forests with large trees, savannas, grasslands, cactus-dominated communities, and other associations; some of natural origin, others anthropogenic. The species-rich mesic forests occupy low and high elevation slopes. On windward-facing slopes of northeastern Hispaniola, where rainfall is more abundant, the lower boundary of mesic vegetation descends close to the sea, but on leeward slopes it remains at higher elevation where humidity is greater.

Pinus occidentalis is the only pine species presently growing in Hispaniola (Holdridge, 1947). Pine is widely distributed throughout the island and, although a predominantly high mountain species, it also grows near sea level. At $\approx 1,800$ m of elevation and higher, dense, pure stands of large pine trees form extensive open forests with shrubby understories. Just below this altitudinal level, pine-hardwood forest associations are commonly established in sites with rich soils and low incidence of fires. At lower elevations, pine populations are scattered along ravines or found on poor soil sites, and have fewer trees than summit forests. Because pines are very resistant to fire, except in the juvenile stage, they can form pure stands in dry areas with frequent fires (Holdridge, 1945, 1947).

In Haiti's southwestern peninsula, where Lake Miragoane is located, the tropical dry forest zone lies along the narrow coastal lowlands, whereas mesic and pine forests are restricted to the slopes and summits of massifs de La Selle and La Hotte. A transect of dry forest, ≈ 150 km in length, separates the mesic communities of the massifs (Fig. 1).

Lake Miragoane's watershed and surrounding areas supported lowland tropical dry forest in the past, but the land has been largely deforested for agriculture and timber exploitation. The drainage basin is presently characterized by agricultural fields in use and abandoned, sparse secondary vegetation, and heavily eroded soils (Higuera-Gundy, 1989). On the slopes facing the south shore of the lake, severe soil erosion has exposed the bedrock. There are a few small remnants of secondary forest with Cecropia, Trema, Bursera, Catalpa, Prosopis, Acacia, and other taxa (J.D. Skean, personal comm.). Otherwise, the bulk of the local vegetation consists of shrubs, weeds, and some cultivated fruit trees such as mango, avocado, coconut palm, etc. (Brenner and Binford, 1988).

The most densely vegetated area near the lake is the large marsh on the east-northeast side of the basin. Cladium jamaicense, a sedge, is the main marsh component. Other submerged or emergent macrophytes found in the marsh include Potamogeton, Vallisneria, Paspalidium, Eleocharis, Najas, Sagittaria, and the alga Chara (J.D. Skean, personal comm.). Nelumbo lutea forms large, thick patches along the edge of the southwestern bay, opposite the marsh.

Culture

The earliest known record of human arrival in the West Indies dates from ≈ 7,000 B.P. (Cruxent and Rouse, 1969; Rouse, 1989). About five millennia later,

agriculture and the art of making pottery were introduced into the islands from South America by the ancestors of the Tainos (Rouse, 1989). The Tainos made ceramics and practiced shifting agriculture for the cultivation of cassava (<u>Manihot</u>), the staple food, as well as corn, beans, potato, arrowroot, and other crops (Rouse, 1948). In Hispaniola, three ceramic series are recognized: Ostionoid, Meillacoid, and Chicoid (Rouse and Moore, 1984).

Migration of the agriculturist Tainos into Haiti's southwestern peninsula has been traced through their pottery, much of which remains undated. Meillacoid ceramics (1,100 - 500 B.P.) are widespread throughout peninsular areas, while those of Chicoid age (800 - 500 B.P.) are far less common and mostly found to the east of Miragoane (Rouse and Moore, 1984). Existence of Ostionoid ceramics, 1,400 - 1,100 B.P., in the peninsula is still uncertain. Near Lake Miragoane, a few Meillacoid and Chicoid sites have been discovered (C. Moore, personal comm.). The sites are located in low elevation hills (280 - 440 m) about 1 km south of the lake shore. Other sites occur farther from the lake within a 5 km radius, but may or may not lie within the watershed.

BACKGROUND RESEARCH

Since paleolimnological research relies on multiple disciplines, this work, whose main scope is pollen analysis, also includes information on the geochemical and isotopic ($^{18}O/^{16}O$) records of Lake Miragoane, Haiti. The chemical and isotopic analyses, as well as the isotopic dating of the sediments, were performed by other members of the Miragoane research project. My colleagues' results, most of them already published, are discussed in conjunction with the palynological record of vegetational change in order to present a more comprehensive ecological history of Lake Miragoane and its watershed. In this section, I report the general methods and/or results of analyses carried out by my co-workers.

Coring of the Lake

The 7.67-m core studied in this work was raised from the central, deepest (41 m) part of Lake Miragoane, Haiti. Coring of the lake was done by Michael W. Binford, Mark Brenner, and James D. Skean, in the summer of 1985. The core comprises two sections: 1) a 72-cm mud-water interface core, obtained with a 4cm-diameter piston corer, and 2) a long core section extending from 1.1 to 7.67 m below the mud-water interface. The long core was raised with a square-rod piston corer (Wright et al., 1984). There is a 38-cm hiatus between 0.72 and 1.1 m of the core. The short core (designated 17- VII-85-1) was extruded and sampled in the field and the longer sediment section in the laboratory. Sediment subsamples were taken for palynological and chemical analyses, and isotopic dating. Shells of <u>Candona</u> sp. were later collected from the core for isotopic analyses.

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Relative Composition of the Sediments

Chemical analyses of Miragoane sediments were carried out by Dr. Mark Brenner. He provided the data on organic matter content (L.O.I. 550°C) and sediment dry density (g/cm^3; Fig. 2), and the concentration values (mg/g) of Ca, Mg, and Fe, which I used for computing $CaCO_3$, $MgCO_3$, and Fe_2O_3 (as equivalents of Ca, Mg, and Fe, respectively). Silica (as SiO_2) was assumed to represent the proportion of the matrix not accounted for by organic matter, carbonates, and iron oxide. Dry density is the dry weight (g) of 1 cc of wet sediment. Most of the values of chemical components reported in the text were computed by the author of this work using Dr. Brenner's original data.

The physical and chemical characteristics of Miragoane lacustrine deposits are shown in Figure 2. For comparison purposes Figure 2, like all other Figures showing results from the Miragoane core, includes the core chronology and seven pollen stratigraphic zones. In general, the lacustrine deposits of the long core (1.1 - 7.6 m) are geochemically fairly uniform and lack the the stratigraphic zonation shown by pollen (Figs. 3, 4). In contrast, the sediments of the mud-water interface core (72 -0 cm) are richer in organic matter and display a well-defined stratigraphy determined by chemical and palynological changes at 30 cm (Figs. 2, 3, 4; Table 1; see also Binford et al., 1987; Brenner and Binford, 1988; Higuera-Gundy, 1989). Table 1 shows the average percentage of each major matrix component and the mean dry density for the long core, and each of the stratigraphic sections of the short core.

Figure 2. Proximate composition (%) and dry mass (g/cm^3) of Lake Miragoane deposits. Organic matter is the weight loss on ignition at 550°C. Carbonates represent the sum of $CaCO_3$ and $MgCO_3$ equivalents of Ca and Mg, and iron oxide (Fe_2O_3) is the equivalents of Fe concentrations. Silica, as SiO_2, is the balance of the dry matrix and probably represents detrital silts and clays as well as some biogenic silica. Sediment dry mass is the dry weight of 1 cc of wet sediment oven-dried at 110°C for 24 hours. Note depth scale change.

LAKE MIRAGOANE, HAITI
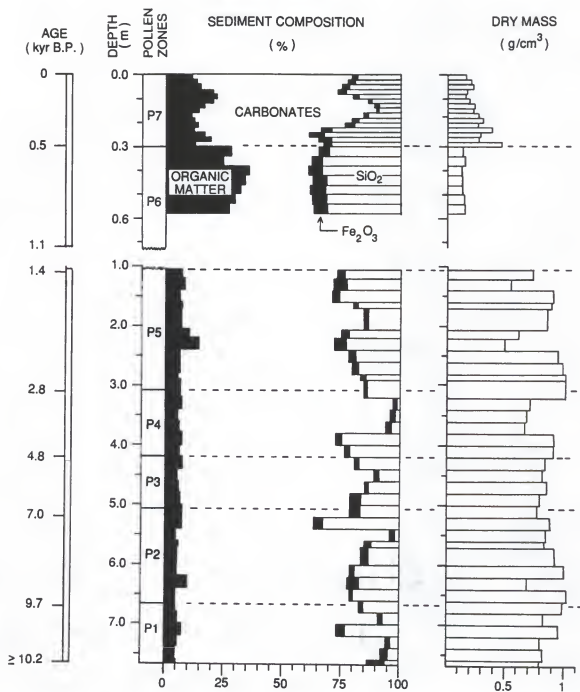
Table 1.- Sediment proximate composition (mean percentages).

Matrix components	Long core 7.6-1.1 m	Mud-water interface core[*] 72-0 cm	
		72-30 cm	30-0 cm
Organic matter	6.4	30.0	15.0
$CaCO_3$	73.0	30.0	58.0
$MgCO_3$	2.6	2.7	3.0
SiO_2	15.5	32.0	18.5
Fe_2O_3	2.6	5.1	3.2
Mean dry mass (g/cm^3)	0.84	0.14	0.25

[*]Some of the mean values for this core are from published results (see text for references).

<u>Isotopic</u> <u>Analysis</u> ($^{18}O/^{16}O$)

Isotopic analysis were performed on shells of <u>Candona</u> sp., an ostracod, collected from 515 levels of the Miragoane core. Collection of shell samples and the isotopic analyses were done by Dr. D.A. Hodell and one of his graduate students, J.H. Curtis, at the Geology Department, University of Florida, Gainesville. The isotopic profile (Fig. 9) is a modified version of the one reported by Hodell et al. (1991 in review). The results are expressed as $\delta^{18}O$, the per mil ($^O/_{oo}$) deviation of each sample's $^{18}O/^{16}O$ ratio with respect to that of the PDB standard. The minimum and maximum isotopic values for the whole core are - $0.42^O/_{oo}$ at

5.13 m and 2.46$^o/_{oo}$ at 7.55 m, respectively. Changes in $\delta^{18}O$ along the core, have been interpreted as indicating large variations in evaporation/precipitation ratios, and major fluctuations in lake level (Hodell et al., 1989, 1991 in review).

<u>Isotopic Dating of the Sediments</u>

The topmost 8 cm of the mud-water interface core (17-VII-85-1) were dated twice by ^{210}Pb assay. Dr. Michael W. Binford ran the procedure. The age at 8 cm is 133 years. Radiocarbon dating of sediments has been performed only on samples from the long core (1.1 - 7.6 m). Dr. Binford submitted a set of bulk sediment samples to the Radiocarbon Laboratory, University of Pittsburgh, Pennsylvania, for conventional ^{14}C dating. Dr. R. Stuckenrath dated those samples. Because several of the dates appeared to be erroneous (e.g. old-age deposits interbedded with sediments up to several millennia younger), a set of ostracod shells were submitted for AMS ^{14}C dating in 1990. Dr. D.A. Hodell collected the ostracod shell samples and sent them to the Woods Hole Oceanographic Institute in Massachusetts, where Dr. G.A. Jones did the AMS dating. Of the AMS dates available in 1990, only those corresponding to core levels documenting major environmental change were used to construct the chronology of the core presented in the results section of this dissertation.

PALYNOLOGICAL TECHNIQUE

Volumetric samples (1 cc) for pollen analysis were processed using a standard technique (Whitehead, 1981). The main steps include treatment of the sediments with diluted solutions of KOH for deflocculation and removal of humic acids, HCl for the removal of carbonates, and HF (40%) for the dissolution of silicates. The samples were acetolized (using a 1:9 mixture of H_2SO_4 and acetic anhydride) to remove excess cellulose and then dehydrated with absolute ethyl alcohol and Tertiary Butyl Alcohol (TBA).

The residue, consisting of pollen and non-oxidized organic matter, was suspended in a known volume of TBA (5 or 10 ml). At the time of slide preparation, the TBA-pollen suspension was homogenized with a magnetic stirrer to ensure an even distribution of particles. One or more aliquots, 0.02 ml each, were placed on a microslide containing a drop of warm silicone oil as mounting medium. Once the TBA evaporated, a cover slip was placed and sealed.

Magnifications of 200X and 300X were used to observe and count pollen and charcoal fragments. Oil immersion microscopy was used as needed for identification of grains. Carbonized plant fragments were counted, but their sizes and potential sources (e.g. from monocots or dicots) were not determined.

A pollen reference collection, which includes numerous taxa of the American tropics, was used to identify the fossil pollen grains, mainly by morphological comparison. Pollen samples of taxa known to occur in Hispaniola were obtained from the herbarium of the University of Florida, and, after processing, the slides were added to the pollen reference collection.

Total pollen counts were brought to 300 grains and included arboreal and non-arboreal taxa, both common and rare, and unidentified grains. Pollen of aquatic plants is included in the total pollen sum while spores of ferns, mosses, fungi, etc. are excluded from this count. The relative abundance of each taxon or group discussed in this work, is with respect to the total pollen sum.

RESULTS

Pollen Assemblage: Percentages and Taxonomic Composition

In this and subsequent discussions, the short mud-water interface core and the longer sediment column are treated as a single profile extending from 0 to 7.6 m, with a 38-cm hiatus between 0.72 and 1.1 m. Pollen percentages for major taxa or groups are presented in Figures 3 and 4. Percentage ranges in this section refer to minimum and maximum values of a taxon or group over the entire core. Unidentified grains (15 observed types) and poorly represented taxa are excluded from all diagrams, though they were included in the total pollen sum of 300 grains. The following list enumerates rare but known taxa. The total number of grains observed along the whole core is given in parentheses: Rhizophora (15), Alternanthera (4), Byrsonima (6), Bocconia (1), Cactaceae (3), Tiliaceae (1), Adelia (3), Juglandaceae (2), Malvaceae (1). When present, each taxon equals less than 1% at any core level. Pollen of Cissampelos, a vine, has a maximum of only 2% but is continuously present between 4.3 and 3.2 m and absent at other levels.

Arboreal pollen of Urticales predominates in the record. Combined proportions of Cecropia, Moraceae, Trema, and Celtis account for about 40% to 80% of all pollen at each sampled level. The percentage of each taxon, however, shows marked fluctuations along the sediment profile. Moraceae is represented by Chlorophora and Trophis pollen types, by Pseudolmedia (presumably P. spuria), and a few grains of Ficus. Additional arboreal elements include Podocarpus, Pinus, and several poorly represented members of dry and moist forests.

Assignment of taxa to the dry and moist forest groups was based on information from local floristic reports describing their general habitat preferences (Judd, 1987; Holdridge, 1947, 1945; Fisher-Meerow and Judd, 1989; Ekman, 1928; Howard, 1973). Among the moist forest group, pollen of Zanthoxylum and Alchornea are abundant and Nectandra is rare. Moist forest pollen amounts to 0.3% to 7.5% of the total pollen sum. The sediments are comparatively richer in pollen derived from the dry forest, 2% to 19% (Fig. 3).

Dry forest representatives include Curatella (probably C. americana), Phyllostylon, Bursera, Spondias, Ampelocera, Guazuma, and several legumes (e.g. Prosopis, Caesalpinia, Mimosa, Cassia, etc.). Grains of Meliaceae (possibly Swietenia), Bombacaceae, and Cordia are infrequent. Pollen of Myrtaceae, Sapotaceae, and Sapindus is more abundant in sediments containing higher representation of dry forest trees and for this reason they were included in this group. However, some species of Sapotaceae and Myrtaceae also grow in moist forests (Ewel and Madriz, 1968).

Pine pollen preserved in deep, older deposits is morphologically similar to that observed in recent sediments. The similarity suggests that all fossilized pine pollen found in Miragoane's deposits belongs to Pinus occidentalis, the only pine species presently growing in Hispaniola (Holdridge, 1945, 1947). Pine pollen representation ranges from 0.3 to 9%, with its maximum value attained in surface muds. Pollen of Podocarpus (0 - 1.6%), Palmae, and the few grains of Rhizophora and Alternanthera are restricted to levels below 5.1 m. The last two genera are not shown in the Figures.

Non-arboreal pollen in the sediments derives from Palmae, shrubs, weeds, and aquatic plants (Fig. 4). Palmae pollen comprises 0 to 10.5% of the sum and is represented by three unidentified types, two common and one rare. The shrub

group is represented by pollen of Melastomataceae (<u>Miconia</u> and another pollen type resembling <u>Conostegia</u>), several members of the Euphorbiaceae (but mainly a type presumed to be <u>Gymnanthes</u>) and, occasionally <u>Myrica</u> and Anacardiaceae. The mean proportion of <u>Miconia</u> along the core is 6.6%, which is twice as high as that of <u>Gymnanthes</u>. <u>Conostegia</u> ($\bar{X}$ = 1.4%) is more common below 5.1 m and quite infrequent in overlying deposits.

The most common weeds are grasses, <u>Ambrosia</u>, <u>Pilea</u> and members of the Chenopodiaceae/Amaranthaceae (hereafter referred to as Cheno/Ams.) which may include terrestrial and/or aquatic representatives at different levels in the core. Uncommon weeds include <u>Acalypha</u>, Cyperaceae, Umbelliferae, Compositae (other than <u>Ambrosia</u>), and Solanaceae, all grouped under the "other herbs" category in all pollen diagrams. Total relative abundance of all herbs, including the Cheno/Ams., varies from 0.3% to 53% along the core. A few <u>Zea</u> pollen grains were seen at 72 cm and in surface muds (Fig. 4).

Littoral vegetation is primarily represented by pollen of <u>Cladium</u> (Cyperaceae) and <u>Typha</u>. <u>Cladium jamaicense</u> and <u>Typha domingensis</u> are presently common in an extensive marsh east of Lake Miragoane. <u>Sagittaria lancifolia</u>, <u>Nelumbo lutea</u>, and <u>Nymphaea ampla</u> also abound in the littoral zone but their pollen is rare in the sediments. In the topmost 72 cm of the core, pollen of aquatics varies from 3.5% to 17.5%; deeper sediments contain 3.5% or less. Pollen of the submerged macrophyte <u>Potamogeton</u>, occurs intermittently through most of the core and is absent in many levels below 1.1 m. Like other aquatics, <u>Potamogeton</u> becomes abundant above 72 cm, reaching its maximum (7.5%) in the top few centimeters of sediment. Values of aquatics in all diagrams exclude the submerged macrophyte.

Figure 3. Pollen percentages of major arboreal taxa or groups of taxa (e.g. moist forest, dry forest, shrubs, etc.). In this and all other Figures the Moraceae includes Pseudolmedia, Chlorophora-T, Trophis-T and a few grains of Ficus. Moist forest: mainly Zanthoxyllum and Alchornea. Dry forest: Curatella, Bursera, Spondias, Phyllostylon, Ampelocera, Sapindus, Guazuma, Myrtaceae, Meliaceae, Sapotaceae, legumes, etc. Note depth scale change.

LAKE MIRAGOANE, HAITI

ARBOREAL POLLEN (%)

BOLD BARS = HALF VALUES
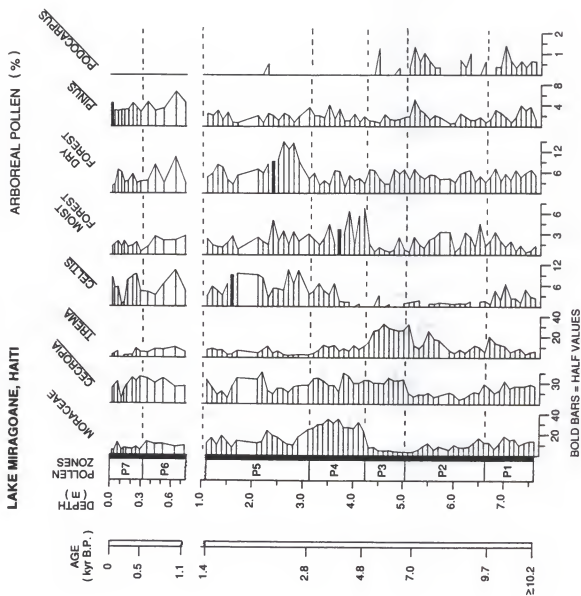
Figure 4. Pollen Percentage of non-arboreal taxa or groups. The shrubs include Miconia, Conostegia-T and Gymnanthes-T. The "other herbs" include various members of the Cyperaceae, Solanaceae, Composites (other than Ambrosia) and other weeds. The aquatics include mostly pollen of Cladium, Typha, some Nelumbo, and less common taxa. Potamogeton is excluded from the values of aquatics in all diagrams. Note depth scale change between cores.

LAKE MIRAGOANE, HAITI

NON-ARBOREAL POLLEN (%)

* ZEA POLLEN PRESENT

BOLD BARS = HALF VALUES
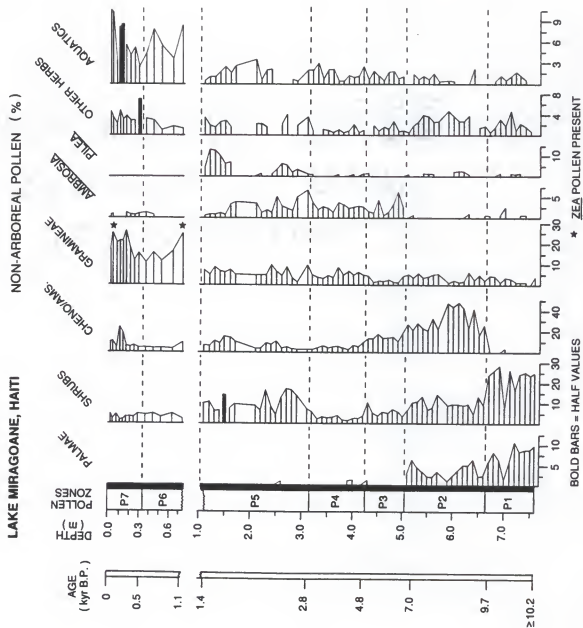
Chronology of the Core

Isotopic Dates

The chronology of the core (Table 2) includes several isotopic dates determined and published by other members of the Miragoane research project. As some levels of the core containing evidence of major environmental change remained undated, I estimated their age through different approaches.

The age of 500 years for the 30 cm level of the mud-water interface core was determined by downcore extrapolation of ^{210}Pb dates obtained for the uppermost 8 cm of sediment. The age at 8 cm is 133 years. Detailed ^{210}Pb dates of the Miragoane mud-water interface core are presented in Brenner and Binford (1988). Based on ^{14}C determination on bulk organic sediment, the age at 7.53 m is 10,230 $\pm$ 160 B.P. and that at 2.2 m is 2,115 $\pm$ 55 B.P. The four AMS dates reported in Table 2 are according to Hodell et al. (1991 in review), and are uncorrected for hard-water error.

Radiocarbon dating of materials (e.g. shells, plants, marl, etc.) developed within hard-water lakes underlain by limestone, like Lake Miragoane, can yield ages up to 2,000 years too old, due to the incorporation of bicarbonate derived from the dissolution of ^{14}C-depleted ("old") limestone (Deevey et al., 1954; Deevey and Stuiver, 1964).

Estimated Dates

In the mud-water interface core, the sediments between 30 and 72 cm were deposited during settlement of the drainage basin by early agriculturists: the Tainos. Archaeological data document the existence of several Taino-age (1,100 - 500 B.P.) sites within the lake's watershed (Rouse and Moore, 1984; C. Moore, personal comm.). Although presence of Zea pollen at 72 cm (Fig. 4) gives evidence of near-

shore agriculture, it cannot be ascertained whether these deposits are contemporaneous with or postdate the earliest known agricultural settlements. Nevertheless, the age estimate of 1,100 B.P. was tentatively assigned to the 72 cm level, pending corroboration by radiocarbon dating.

According to pollen data, the high relative abundance of Moraceae pollen between 4.2 and 3.1 m (Fig. 3), records Hispaniola's most mesic Holocene episode. Based on our chronology, the onset of this mesic climate, $\approx$ 4,800 B.P., was contemporaneous with the demise of coral reefs (shortly after 4,760 B.P.) growing in the Enriquillo Valley, Dominican Republic since the early Holocene (Taylor et al., 1985; Mann et al., 1984). Sediments overlying the coral reef contain fossil bivalves indicative of a shift from marine to brackish water. This brackish environment lasted until about 2,820 B.P. Since then, the water of Lake Enriquillo, located within the Enriquillo Valley, has become more saline due to high evaporation (Taylor et al., 1985). The authors partly attribute the shift to brackish conditions to increased fresh water inputs from Rio Yaque del Sur into the saline valley. The river is located to the east of Lake Enriquillo. It is suspected that abundant rainfall, documented by the pollen record of the Moraceae zone, was largely responsible for enhancing fresh water deliveries to the Enriquillo Valley and for the eventual development of the brackish environment at that site.

The nature and initial timing of the environmental changes recorded by the fossil records of Miragoane and Enriquillo during the middle Holocene clearly suggest a common response to the establishment of wet conditions. Like its inception, the end of the moist climate might have been closely synchronous at both sites. Under this assumption, the 3.1 m level of the Miragoane core, marking the transition from wet to dry conditions, was given an age of 2,800 B.P.

TABLE 2.- Chronology of the Core (0 - 7.6 m).

| | | | Sediment Accumulation Rates | |
Depth (m)	Age (years)	Method	Interval (m)	Rate (g/cm^2/yr)
0.08	133	^{210}Pb[**]	0.00 - 0.08	0.010
0.30	500	^{210}Pb[**]	0.00 - 0.30	0.015
0.72	1,100	Estimated age[&]	0.30 - 0.72	0.010
1.10	1,360	Estimated age[*]	0.72 - 1.10	core hiatus
2.20	2,115 ± 55	^{14}C[#]	1.10 - 2.20	0.110
3.10	2,800	Estimated age[@]	2.20 - 3.10	0.115
4.18	4,780 ± 60	AMS ^{14}C	3.10 - 4.20	0.044
5.20	6,945 ± 65	AMS ^{14}C	4.20 - 5.20	0.040
6.22	9,005 ± 75	AMS ^{14}C	5.20 - 6.20	0.044
6.71	9,700 ± 90	AMS ^{14}C	6.20 - 6.70	0.068
7.53	10,230 ± 160	^{14}C[#]	7.60 - 6.70	Undetermined

[**] Age of the uppermost 8 cm of sediment was determined by ^{210}Pb assay, and that at 30 cm by downcore extrapolation

[#] Dates obtained by radiocarbon dating of bulk organic sediments.

[&] Date based on local archaeological chronologies

[@] Age estimate based on chronological data from Lake Enriquillo, Dominican Republic

[*] Age estimated assuming constant sedimentation between 2.2 and 0.72 m.

AMS ^{14}C determination on ostracod shells

See text for details and references.

An age of 1,360 B.P. was estimated for the 1.1 m core level by assuming constant rate of sedimentation between 2.2 m (^{14}C = 2,115 B.P.) and 72 cm (1,100 B.P.). Pollen and charcoal concentrations for the 3.1 to 1.1 m interval (Fig. 5) indicate great fluctuations in the mean rate of sediment deposition calculated for this core section. Thus, the age estimated for the 1.1 m level may be younger or older than the current estimate suggests.

Figure 5. Pollen zones' chronology, core depth boundaries, and mean
sedimentation rates ($g/cm^2/yr$). Microfossil concentration profiles show the
level-to-level values (in $\#/cm^3$ of wet sediment) of total pollen (known and
unidentified) and charcoal. Note depth scale change.

Recent dating of ostracod shells from the 7.18 m level gave an age of 10,300 ± 80 B.P. (D.A. Hodell, personal comm.), thus suggesting that the basal sediments (7.53 m) may be older than 10,230 B.P., the age previously determined by dating bulk organic carbon. For this reason, in all Figures and in the text, the age of the basal deposits is written as ≥ 10,230 B.P. The new AMS date is not included in Table 2 nor was it used to calculate accumulation rates of any parameters.

Pollen-Defined Stratigraphy

Seven stratigraphic zones are defined by major changes in the relative abundance (Figs. 3, 4) and accumulation rates (Figs. 6, 7) of major taxa. Figure 5 shows the chronology, depth boundaries, and the mean sedimentation rate for each zone. Only those dates corresponding to transitional episodes of environmental change are shown in the Figures, though others are included in Table 2. For clarity, some dates presented in Table 2 have been rounded off to the nearest century (e.g. 1,360 = 1,400, 4,780 = 4,800, etc.), except the one at 6.7 m (6,945 B.P.) which was converted to 7,000 B.P. These modified dates appear throughout the text and in all Figures showing the core chronology, though the rates of accumulation of all variables were calculated according to the original dates given in Table 2.

Total pollen and charcoal concentrations (in $\#/cm^3$ of wet sediment) are also included in Figure 5. The wide variations in microfossil concentrations over the core are attributable to several factors such as post-depositional diagenesis, changes in supply, and variable sedimentation. In general, the similarity in total pollen and charcoal concentration profiles suggests control by changes in bulk sedimentation

LAKE MIRAGOANE, HAITI

SEDIMENTATION RATE (g/cm^2/yr)

MICROFOSSIL CONCENTRATIONS (#/cm^3)

POLLEN (–·–·–·–)
CHARCOAL (———)

AGE (kyr B.P.)

DEPTH (m)

POLLEN ZONES

• ≤ 1000/cm^3
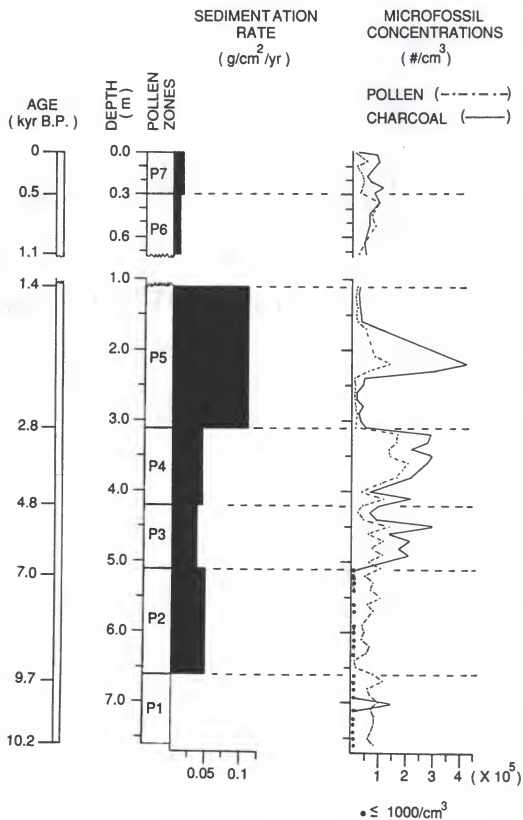
Figure 6. Pollen accumulation rates (#/cm^2/yr) of major arboreal taxa or groups. Values for zone P1 (7.6 - 6.6 m) are not included because the age of the basal sediments is pending verification. Note depth scale change.

LAKE MIRAGOANE, HAITI

ARBOREAL POLLEN (grains/cm² /yr x 10²)

PODOCARPUS · PINUS · DRY FOREST · MOIST FOREST · CELTIS · TREMA · CECROPIA · MORACEAE

POLLEN ZONES · DEPTH (m) · AGE (kyr B.P.)

BOLD BARS = HALF VALUES
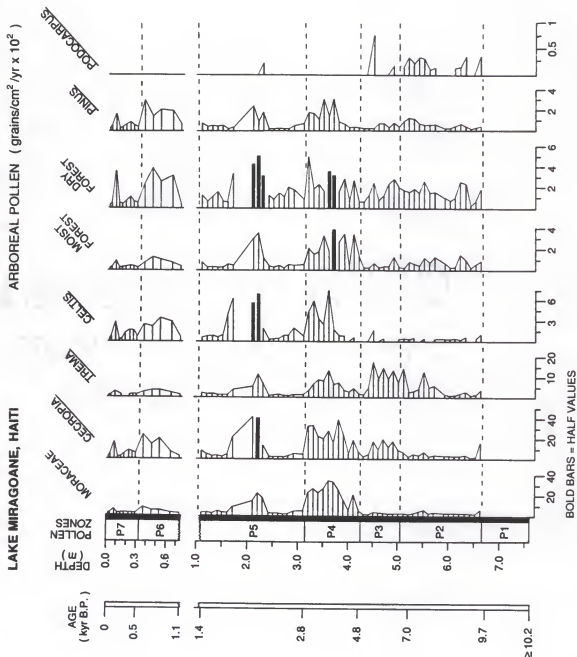
Figure 7. Pollen accumulation rates (#/cm^2/yr) of non- arboreal taxa or groups of taxa. Values for zone P1(7.6 - 6.6 m) are as yet undetermined. Note depth scale change between cores.

LAKE MIRAGOANE, HAITI   NON-ARBOREAL POLLEN (grains/cm² /yr x 10²)
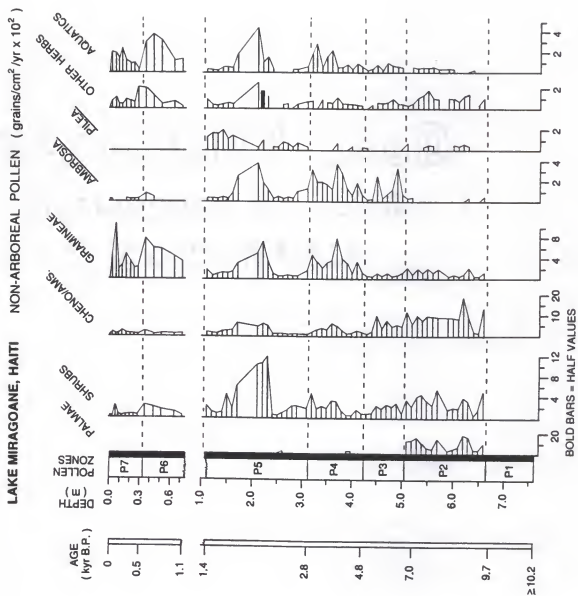
Figure 8. Mean zonal accumulation rates of individual taxa or groups, arboreal and non-arboreal pollen, charcoal fragments (all expressed as #/cm^2/yr), and chemical components, expressed in mg/cm^2/yr. Weed pollen includes grasses, _Ambrosia_, _Pilea_ and all the taxa comprising the "other herbs" group. The arboreal pollen profile includes all the Urticales and the dry and moist forest taxa. Excluded from these values are _Podocarpus_, _Pinus_, and rare taxa. The non-arboreal curve comprises pollen of shrubs, Cheno/Ams., and all terrestrial weeds, including taxa of the "other herbs" group. Pollen of Palmae, aquatic plants, and poorly represented taxa are excluded. Core depth is not to scale.

MEAN ZONAL ACCUMULATION RATES

LAKE MIRAGOANE, HAITI
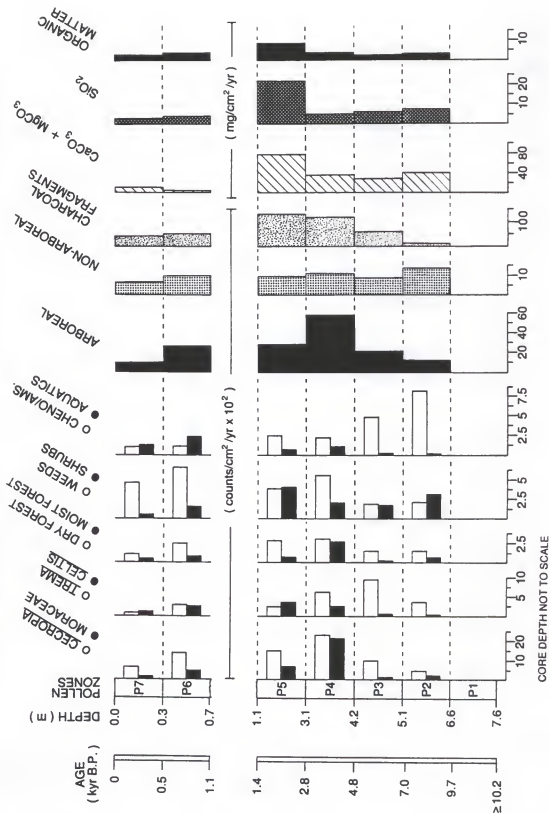
rate. The influence of this and other factors in controlling microfossil concentrations in the sediments is discussed in later sections.

Using a quantitative technique for palynological and chemical analyses, allowed calculation of the concentrations of pollen, charcoal, and chemical components. Their accumulation rates were computed once the chronology of the core was determined.

Mean zonal accumulation rates of arboreal and non- arboreal pollen, individual main taxa or groups, chemical components, and charcoal within each zone, are shown in Figure 8. The mean accumulation values of variables within each zone (usually given in parentheses) are used to define a zone, or to establish comparisons between zones, though level-to-level changes are also cited. Only maximum and minimum oxygen isotopic values (not averages) are cited for each zone.

P1: Shrubs-Palmae-Podocarpus Assemblage Zone

Zone P1: ≥ 10,230 - 9,700 B.P.; 7.6 - 6.6 m. Because an AMS date of 10,300 B.P. at 7.18 m suggests an older (than 10,230 B.P.) age for the basal sediments of the core, whose redating is still pending, the accumulation rates of pollen, charcoal, and chemical components were not determined for zone P1. For this reason, percentage and/or concentration values are used to discuss the characteristics of this period.

The non-arboreal community of this episode is mainly represented by pollen of shrubs and Palmae (Fig. 4), each averaging 22% and 7%, respectively. The bulk of shrub pollen consists of 3 - 19% Miconia, up to 7% Conostegia, and 2.5% Gymnanthes. The proportion of pollen from terrestrial and aquatic herbs is quite low. Charcoal fragments maintain very low concentrations in this zone (usually ≤ 1,000/cm^3), but briefly rise to 140,000/cm^3 at 7.0 m (Fig. 5).

Miconia is represented in Hispaniola by 64 species (Howard, 1973). About 18 Miconia species have been reported from the massifs de La Hotte and La Selle in southern Haiti, where most of them grow at 1,100 m of elevation and higher and vary in abundance from rare to very common (Judd, 1987; Holdridge, 1947; Fisher-Meerow and Judd, 1989). Shrubs of Conostegia icosandra (occasional) and Gymnanthes lucida (common) also occur at similar altitudes in the Massif de la Hotte (Judd, 1987). Conostegia icosandra is common in the seral vegetation of the rainforest of Dominica, as are several Miconia species which also occur in other forests (Hodge, 1954). Despite the morphological resemblance between the fossil grains identified as Conostegia and Gymnanthes and the pollen from their living counterparts, it cannot be determined if the fossil grains belong to the species just mentioned above.

In zone P1, Cecropia has the highest relative abundance (Fig. 3) of all trees with an average of 23%, which is twice as high as that of Moraceae and at least five times greater than any other arboreal group or taxon except Trema, whose zonal mean is 9%. Members of the dry forest present in this episode include Myrtaceae, Sapotaceae, Curatella, Ampelocera, Guazuma, and legumes. The moist forest is represented by Zanthoxylum and, occasionally, by Alchornea. Celtis was common in the area ($\overline{X}$ = 3%). Podocarpus is much less abundant ($\overline{X}$ = 0.5%) but occurs throughout the zone, and Rhizophora, though present, is rare. The Moraceae is represented by Trophis-T and Chlorophora-T through the entire zone, but at 7.0 m the family is also represented by pollen of Pseudolmedia, which amounts to one third of the family's 17% representation at this level. Otherwise, Pseudolmedia is absent between 5.2 m (top of zone P2) and the bottom of the core.

Like all core sediments below 1.1 m, those of P1 have high dry density values (Fig. 2), and in this zone they average 0.86 g/cm^3. On the average, the matrix

consists of 80% CaCO$_3$, 10% silica, 5.4% organic matter, and 2.3% iron oxide (see Fig. 2 for level-to-level percentage values of these components). MgCO$_3$ ($\overline{x}$ = 2.3% in P1) is plotted with CaCO$_3$ as total carbonates in the Figure.

This core section contains some of the highest δ^{18}O values reported from Lake Miragoane (Fig. 9). Most values are positive (0.1 to 2.46^o/$_{oo}$), except for a negative one recorded at 7.03 m (- 0.02^o/$_{oo}$; D.A. Hodell personal comm.). δ^{18}O values higher than 1 per mil characterize the sediments below 6.67 m.

P2: Chenopodiaceae/Amaranthaceae Assemblage Zone

Zone P2: 9,700 - 7,000 B.P.; 6.6 - 5.1 m. Pollen grains of Chenopodiaceae/Amaranthaceae (the Cheno/Ams. group), virtually absent in P1, become predominant in zone P2 (Figs. 4, 7) where they attain an average net accumulation rate of 830/cm^2/yr (Fig. 8). The shrubs-Palmae-Podocarpus assemblage persists throughout this interval, with each component remaining well represented. Nevertheless, in zone P2 the average percentage of shrub pollen (14.4%) and Palmae (4.4%) has declined with respect to the values recorded in the underlying zone. Accumulation of shrub pollen in P2 averages 310/cm^2/yr (Fig. 8) and that of Palmae 85/cm^2/yr (not shown). Alternanthera, restricted to this zone, and Rhizophora are infrequent. The most abundant weeds in this episode are grasses and sedges (Fig. 4; pollen of Cyperaceae makes up the bulk of the "other herbs" group in zone P2). Their combined mean accumulation reaches nearly 200/cm^2/yr in this zone (Fig. 8). Charcoal particles accumulated at a slow mean rate, 150/cm^2/yr, in this core section (Fig. 8).

Dry and moist forest pollen accumulations begin to increase in the bottom part of the zone, and Cecropia and Trema increase in the upper half (above 6.0 m;

Fig. 6), though the latter genus shows a much greater increment. Of the dry forest components, Ampelocera and Guazuma become sporadic throughout the zone.

Bulk sediment accumulation averages 0.05 g/cm^2/yr in this zone (Fig. 5). Mean deposition rates, in mg/cm^2/yr, of matrix components are as follows: total carbonates 39.4, silica 7.2, organic matter 2.8 (Fig. 8), and iron oxide 1.2 (not shown). Average percentages of these components in P2 are comparable to those observed in the previous zone, with the exception of silica, which increases from 10% in P1 to 14.2% in P2.

The range of δ^{18}O values within zone P2 varies between - 0.42 to 1.48^o/$_{oo}$ (D.A. Hodell, personal comm.). A short-term decrease in δ^{18}O recorded between 6.3 and 6.1 m, is followed by a sharp increase, which ends at about 6.0 m. From this level to the end of the zone (5.1 m), δ^{18}O values decrease once again (Fig. 9).

P3: Trema-Cecropia Assemblage Zone

Zone P3: 7,000 - 4,800 B.P.; 5.1 - 4.2 m. Maximum accumulation of Trema (900/cm^2/yr), and accelerated deposition of Cecropia pollen (975/cm^2/yr) characterize this zone (Figs. 6, 8). While the mean accumulation of dry forest pollen remains nearly unchanged, that of the moist forest elements sharply declines (Fig. 8). Occasional occurrence of Podocarpus, Palmae (Figs. 6, 7) and Conostegia are recorded in this zone. Rhizophora ceases to appear at 3.9 m. Major changes in the taxonomic composition of the dry forest when this period began include the near-loss of Myrtaceae and Guazuma coupled with the reappearance of Ampelocera and the introduction of Spondias and Bursera as regular elements of this community. The sediments of P3 show comparatively higher concentrations and accumulation rates of charcoal fragments (5,900/cm^2/yr) than those of zone P2 (Figs. 5, 8).

A drop in mean non-arboreal representation is mostly attributed to a decrease in pollen of shrubs and Cheno/Ams. (Fig. 8). Despite declining, the Cheno/Ams. continue to dominate the non-arboreal pollen rain in P3, with an average of $490/cm^2/yr$. Ambrosia, previously rare, becomes abundant in this zone, surpassing the accumulation values of grasses (Fig. 7). Gymnanthes is lacking in this zone. Total pollen from all terrestrial weeds and the shrubs accumulated at a similar mean rate ($\approx 180/cm^2/yr$; Fig. 8).

In contrast with zone P2, the isotopic record of zone P3 shows less marked fluctuations and all $\delta^{18}O$ values fall within a positive range, 0.08 to $0.73^o/_{oo}$ (Fig. 9).

P4: Moraceae Assemblage Zone

Zone P4: 4,800 - 2,800 B.P.; 4.2 - 3.1 m. Moraceae's highest accumulation rate $(2,100/cm^2/yr)$ is recorded in this interval. The Moraceae rise occurs concurrently with the highest zonal accumulation rates of Cecropia, dry and moist forest trees (Figs. 6, 8), and Pinus (Fig. 6). Accumulation of Trema pollen briefly declines in the bottom of zone P4, and then increases in the upper half, but its representation is never again as high as in zone P3 (Fig. 6). Since the beginning of the Moraceae episode, Celtis occurs regularly in the fossil record maintaining high values through most of the zone (Fig. 6).

Approximately 50% of the dry forest pollen belongs to Spondias and Ampelocera, and the remaining portion includes several less abundant taxa. Pollen inputs from the moist forest derive from two main taxa: Zanthoxylum and Alchornea, each similarly abundant in these deposits. At the transition from P3 to P4, the mean accumulation of pollen from dry forest trees doubles, to $300/cm^2/yr$,

whereas that of the moist forest elements increases almost tenfold, to $265/cm^2/yr$ (Fig. 8).

Throughout the time span covered by P4, Pseudolmedia was a prominent member of the Moraceae forest, often contributing up to 30% of the pollen supplied by this family. In sediments postdating this zone, the representation of this taxon is much lower and frequently discontinuous. Prior to the Moraceae peak, Pseudolmedia is absent at all sampled levels, except for a few grains obsrved at 4.7 and 5.2 m (zone P3), and at 7.0 m (zone P1), where it represents 5% of the pollen sum.

There is a slight increase in the average accumulation rate of non-arboreal pollen in the zone, accounted for by accelerated deposition of weed pollen ($550/cm^2/yr$). Cheno/Ams. rates decrease by half (to $230/cm^2/yr$) whereas inputs from the shrubs show very little change (Fig. 8). Grasses and Ambrosia comprise most of the weed pollen, maintaining an approximate ratio of 2:1 along the zone (Fig. 7). The shrubs are once again represented by Gymnanthes in the Moraceae zone.

Deposition of charcoal fragments increases twofold in zone P4 ($11,500/cm^2/yr$; Fig. 8). Carbonized plant fragments in zones P4 and P3 are large ($\geq 50\,\mu$), well preserved, and include many epidermal remains of monocots. In contrast, those in the two deepest zones of the core (P2 and P1) are much smaller ($\leq 50\,\mu$), worn, and lack morphological features indicative of their plant sources.

$\delta^{18}O$ values in P4 range from 0.1 to $1.2^o/_{oo}$ (D.A. Hodell, personal comm.). The values at 3.6 m and above are nearly twice as high as those recorded in the bottom section of the zone (Fig. 9).

P5: Celtis-Dry Forest Assemblage Zone

Zone P5: 2,800 - 1,400 B.P.; 3.1 - 1.1 m. This episode is characterized by a major decline in the accumulation of arboreal pollen, attributed to drastic reductions in the average deposition rates of Moraceae (by 72%), Cecropia (by 40%), Trema (by 61%), and the moist forest group (by 75%; see Figs. 6, 8). The highest mean accumulation rate of tree pollen within any given zone of the core is recorded in zone P4, $5,700/cm^2/yr$. This value drops to $2,800/cm^2/yr$ in zone P5 (Fig. 8); a net loss of 51%. Unlike all other trees, Celtis increases within zone P5, where it reaches its maximum mean zonal accumulation, $330/cm^2/yr$, equivalent to a 25% increment with respect to zone P4. Although dry forest representation was slightly reduced in P5 (by 1%), this group attains its greatest taxonomic diversity during this episode. Taxa contributing to this taxonomic richness include: Cordia, Phyllostylon, Adelia, pollen of Bombacaceae and Meliaceae (presumably Swietenia). These taxa are either absent or extremely rare in deeper core sections, and in P5 they are less abundant than the major components of this forest such as Curatella, Spondias, Bursera, legumes, and others.

The mean accumulation of non-arboreal pollen in P5 is basically the same as in zone P4, but the average deposition of shrubs increases and that of the herbs decreases within zone P5 (Fig. 8). Representation of Cheno/Ams. stays the same as in P4. There are some major changes in the taxonomic composition of the non-arboreal group (Fig. 7). From 2.4 to 1.6 m Gymnanthes dominates the shrub group; otherwise it is as abundant as Miconia. In descending order of abundance, grasses, Ambrosia, and Pilea are the dominant herbs in the bottom of P5 (below 1.6 m). Pollen of Pilea is mainly confined to this zone but is absent between 2.4 and 1.6 m, except for one occurrence. When present, this genus is as frequent and abundant as

other weeds, although from 1.6 to 1.1 m, it generally contributes more pollen than any other non-arboreal element, including the shrubs.

The decline in forest cover in zone P5 is synchronous with the fastest rate of bulk sedimentation recorded in the core, 0.11 g/cm^2/yr (Table 2, Fig. 5). Net accumulations of all sediment components also reach maximum averages in this zone. Total carbonate accumulation (80.1 mg/cm^2/yr) is about four times greater than silica and nearly an order of magnitude higher than organic sedimentation (Fig. 8). Iron oxide too, reaches maximum rate of deposition in this zone ($\overline{x}$ = 3.5 mg/cm^2/yr; not shown in Fig. 8).

This core section is further characterized by recording three short-term episodes of major, concurrent changes in the concentrations of total pollen and charcoal fragments (Fig. 5). In the bottom of the zone, 3.1 - 2.4 m, pollen averages 10,600/cm^3 and charcoal 30,000/cm^3. Similarly low values are recorded from 1.6 to 1.1 m. In contrast with these two microfossil-depleted sections, the middle interval (2.4 - 1.6 m) is, on the average, richer in pollen and carbonized particles: 56,300/cm^3 and 190,000/cm^3, respectively. The fact that both plant fossils undergo simultaneous changes (in concentration) of comparable magnitude suggests that the average sedimentation rate for zone P5 did not remain constant. One of the assumptions in this work is that sediments accumulated at a constant mean rate over the time span defined by two dated levels in the core. Simultaneous, pronounced changes in pollen and charcoal concentrations suggest that this assumption is violated at times throughout the lake's sedimentary history, but more so within the time encompassed by zone P5.

Fossilized charcoal fragments in zone P5, and in all remaining sediments above this zone, are represented by a mixture of small, worn particles and large fragments of epidermal remains of monocots.

$\delta^{18}O$ values, though sharply increasing at the bottom of the zone, decline between the 2.3 to 1.4 m interval, and then rise again through the rest of the core (Fig. 9).

P6: Taino Episode

Zone P6: 1,100 - 500 B.P.; 72 - 30 cm. The time span encompassed by zones P6 and P7 represents known periods of human occupation of Miragoane's watershed. For this reason, it is more appropriate to name these zones according to the cultural episode each one represents. With only minor taxonomic and numerical changes, the forest characterizing zone P5 persists throughout Taino times. Level-to-level and mean accumulation values of shrubs and Cheno/Ams. pollen decline by more than half while those of herbs and aquatics sharply increase (Figs. 7, 8). Pollen of terrestial weeds, dominated by grasses, and the aquatics reach maximum mean accumulation in P6: 650/cm^2/yr and 240/cm^2/yr, respectively (Fig. 8). Pollen of Acalypha, absent in underlying zones, becomes common during the Taino episode. This genus accounts for up to 50% of the "other herbs" pollen at most levels of zones P6 and P7 (Fig. 4). Zea pollen occurs at 72 cm (Fig. 4). Pilea no longer appears in sediments of the 72-cm core, and Ambrosia ceases to occur in most Taino-age deposits.

Major quantitative changes in geochemistry, rather than pollen, distinguish the Taino zone from zone P5 just below. A set of ratios emphasizes the marked differences between the values recorded in P5 and those of P6. Comparisons of all sediment-related variables are based on mean accumulation rates within each zone, expressed as mg/cm^2/yr. Accumulation rates for P5 divided by corresponding rates for P6, give the following ratios: a) bulk sedimentation, 11:1, b) total carbonates 25:1, c) silica 7:1, d) iron oxide 7:1, e) organic matter 3:1, f) mean dry density in

g/cm^3, 6:1. Ratios of the mean accumulation of plant fossils, represented as #/cm^2/yr, are: g) charcoal fragments 3:1, h) non-arboreal pollen 1:1, i) arboreal pollen 1:1 (Figs. 2, 5, 8).

A major shift in sediment composition, from inorganic in P5 to more organic in P6, is responsible for the drastic differences in the geochemical record of these two zones. Taino deposits are five times more organic ($\overline{X}$ = 30%) and contain only half the proportion of carbonates ($\overline{X}$ = 32.6%) present in zone P5 (Fig. 2). The organic nature of the matrix accounts for its low dry density ($\overline{X}$ = 0.14 g/cm^3) and to a large extent, for the order of magnitude drop in dry mass accumulation rate, from 0.11 g/cm^2/yr in P5 to 0.01 g/cm^2/yr in P6 (Fig. 5). As a result, the deposition rates of chemical components, expressed in terms of dry weight, were sharply reduced. Carbonate accumulation in the Taino zone (3 mg/cm^2/yr) is the lowest recorded in the core. In this episode, the deposition rates of siliceous and organic materials resemble those of total carbonate (Fig. 8).

P7: Post-Columbian Episode

Zone P7: 500 B.P. - Present; 30 - 0 cm. Accumulation rates of all tree taxa and the shrubs fall to minimum values in this episode (Figs. 6, 7). Total mean deposition of grass-dominated weed pollen (450/cm^2/yr; Figs. 7, 8) equals the combined mean representation of all trees excluding Cecropia, the most abundant forest element in this episode. Cecropia alone contributes an average of 550 grains/cm^2/yr (Fig. 8). Pollen of aquatics, although still well represented, has declined to nearly half the values reported for P6 (Fig. 7). Bursera (dominant) and Sapindus (less abundant) constitute the main dry forest trees while all other taxa are rarely present. Pollen of corn is present but uncommon in surface deposits (Fig. 4).

Average bulk sedimentation in Post-Columbian times (0.15 g/cm^2/yr) is about 50% faster than in the Taino period (Fig. 5). Carbonate accumulation triples in the Post- Columbian period (to 9 mg/cm^2/yr), but the deposition rates of silica and organic matter are slightly lower than in P6 (Fig. 8). Due to a twofold increase in carbonate ($\overline{X}$ = 61%) and lower organic matter content ($\overline{X}$ = 15%), the Post-Columbian muds are heavier ($\overline{X}$ = 0.25 g/cm^3) than Taino deposits (Fig. 2, Table 1).

At 7 cm depth, total pollen concentration sharply increases (Fig. 5). The palynological shift closely coincides, stratigraphically, with a geochemical change recorded between 8 and 10 cm. Within this 2-cm interval, organic matter increases to 21.5% and total carbonates decrease to 58% (Fig. 2). A similar trend of organic enrichment and carbonate depletion is observed in two sampled levels immediately above and below the 8 to 10 cm interval. According to ^{210}Pb chronology, the sediments at 8 cm are about 133 years old and those at 7 cm are about a century old (Brenner and Binford, 1988).

VEGETATIONAL HISTORY OF LAKE MIRAGOANE AND ITS CLIMATIC IMPLICATIONS

General

The palynological record of Lake Miragoane, Haiti provides evidence of major vegetational changes in Hispaniola during the late Pleistocene and throughout the Holocene. In addition, isotopic data reveal long-term fluctuations in lake level concurrent with the vegetational shifts (Hodell et al., 1989, 1991 in review). Most of the environmental changes recorded in the terrestrial and lacustrine systems predate settlement of the study site some 1,100 years ago and are, therefore, considered to have been climatically-induced. The timing, magnitude, and nature of these events are comparable to those recorded in other nearby lowland sites, thus suggesting a common response to a regional climate that most likely affected other Caribbean islands as well.

Although poorly reflecting the species-rich flora of the area, the pollen record comprises representatives of several communities growing along an altitudinal gradient, from aquatic associations of the littoral zone to pine forests of the summits, thereby providing an insight into the wide variations experienced by these communities in the past.

Late Pleistocene (P1: > 10,230 - 9,700 B.P.)

This episode is represented by the shrubs-Palmae- <u>Podocarpus</u> assemblage zone (P1: 7.6 - 6.6 m). Since accumulation rates of pollen, carbonized plant fragments, and chemical components have not yet been determined for zone P1,

their respective averages and/or ranges in relative abundance or concentration will be used to discuss the environmental history of this episode.

According to palynological evidence, the watershed was occupied by plant communities dominated by shrubs and palm trees during this episode. Percentages of tree pollen are low (Fig. 3), revealing scarcity of forests in the region. The charcoal record (Fig. 5) indicates rarity of local and regional fires. Isotopic data (Fig. 9) and presence of fresh-water ostracods (Candona sp.) indicate that the lake, though shallow, contained fresh water during this phase.

Mountain slopes around the basin supported shrub communities dominated by species of Miconia, Conostegia, and Gymnanthes associated with small populations of trees such as Trema, Podocarpus, and Zanthoxyllum (Figs. 3, 4). All these taxa, including the shrubs, are currently common at elevations of ≈ 1,100 m and higher and form part of the montane hardwood forest, whose lower boundary begins at ≈ 800 m (Holdridge, 1947, 1945; Judd, 1987; Fisher-Meerow and Judd, 1989). Celtis too, was common at some distance from the lake. Given the large number of Miconia species presently growing in Hispaniola (64 according to Howard, 1973), the species comprising this ancient vegetation probably included elements from upland-moist and lowland-dry habitats. Those from the latter environment may no longer be seen around the basin due to severe local deforestation. The abundance of shrub pollen ($\overline{X}$ = 22%), comparatively higher than any arboreal taxon except Cecropia ($\overline{X}$ = 23%), suggests predominance and ample distribution of the shrub community.

Existence of forests in warm, moist, lowland habitats is documented by Cecropia and Moraceae, a family represented by Trophis and Chlorophora types of pollen. Low percentages of these taxa suggest that during this period, such forests

were very small, mostly successional, quite far from the lake, and perhaps fragmented. Moraceae averages 12% of the total pollen sum in P1.

The palynological record points to aridity in the lowlands and dry conditions in the highlands of Hispaniola at the end of the Pleistocene. Low lake level, inferred from the $\delta^{18}O$ record, reveals a high evaporation/ precipitation ratio for the period (Hodell et al., 1989, 1991 in review). Pollen of palms varies from 5% to 11% within zone P1 except for a low of 1% at 7.0 m (Fig. 4). This unusually high abundance of palm pollen suggests near-shore distribution of these plants. At present, several species of palms occur within Hispaniola's lowland dry communities, and some (e.g. Pseudophoenix vinifera) expand into the ecotone of the dry- very dry forest formations (Woodring et al., 1924; Holdridge, 1945, 1947). Considering the severely dry conditions of the late Pleistocene, it is possible that the palm pollen present in zone P1 (and also in P2) may belong to xeric species. As their abundant pollen suggests proximity to the Miragoane basin, their most likely habitat was probably the lowlands immediately surrounding the lake.

This near-shore open community supported a few tree taxa, namely Ampelocera, Guazuma, Myrtaceae, and some legumes (all included in the dry forest group; Fig. 3). Although grasses were the most common weeds, they were not abundant in this episode (Fig. 4).

Of all the upland taxa represented in the fossil record, the pollen of Podocarpus and Conostegia (a Melastomataceae shrub) is mostly confined to zones P1 and P2 (Figs. 3, 4; Conostegia is plotted with the other shrubs in all Figures). The restricted occurrence of the two montane elements coupled with high abundance of other highland taxa (e.g. Miconia and Gymnanthes) in these zones, could indicate that during the two episodes represented by these zones, the hardwood forest lower boundary extended closer to the lake than it does at present.

Long distance pollen dispersal may partly explain the occurrence of Podocarpus and Conostegia in zones P1 and P2, as open forest conditions enhance the representation of distant taxa in lacustrine deposits. However, the absence of these two taxa in other core sections recording open landscapes (e.g. zones P5 to P7; Figs. 3, 4, 8) lends credence to the hypothesis that these elements lived in closer proximity to the basin at the end of the late Glacial (P1) and during the early Holocene (P2).

Early Holocene (P2; 9,700 - 7,000 B.P.)

The Chenopodiaceae/Amaranthaceae assemblage zone (P2: 6.6 - 5.1 m) represents this phase. The shrubby and xeric communities characterizing the late Pleistocene persisted into the early Holocene (P2). Although slightly more forested than in the previous episode, these communities remained rather open. Lowland moist forests remained small and forest fires continued to be rare. Progressive forestation of the watershed began at the onset of the Holocene, and coincided with rapid filling of the lake. Collectively, this evidence suggests amelioration of the former climate which nonetheless remained drier than the modern regime.

Within this episode, there is a well-defined trend of increased arboreal representation throughout the zone. From 6.6 to 6.0 m however, pollen concentrations (Fig. 5) display sharp fluctuations, with a major decrease recorded at 6.5 and 6.4 m. Low pollen counts/cc and near-absence of ostracod shells in these two levels (D.A. Hodell, personal comm.) strongly suggest rapid sedimentation and consequent dilution of fossils, shortly after the basin filled 9,700 years ago (6.6 m; see isotopic curve in Fig. 9). Based on arboreal pollen accumulation rates (Fig. 6), the evidence of forest growth is more apparent from 6.0 m to the top of the zone at

5.1 m, though the forestation process probably started at the inception of this period.

Dry forest trees, with Curatella dominant, may have grown next to or associated with the xeric palm community. Curatella, along with Byrsonima, is presently very common in West Indian savannas (Holdridge, 1945). Byrsonima pollen however, was rarely observed in the sediment samples. Pollen percentages of legumes, Myrtaceae, and Sapotaceae also increased within zone P2 while Guazuma and Ampelocera became rare. Because many tropical tree species are not wind pollinated (Flenley, 1973), the degree of forestation in the lowlands may be underestimated by the fossil record. Grasses and sedges also became abundant in the area (Fig. 7; sedges comprise most of the "other herbs" pollen in P2).

At the start of the Holocene, mountain slopes near Lake Miragoane were still dominated by shrubs, but the populations of Podocarpus and Zanthoxyllum began to expand their coverage. Within the time span covered by the upper half of zone P2 (above 6.0 m), the accumulation of pollen of Trema, pine, and shrubs increased (Figs. 6, 7), indicating that the highland communities expanded and/or became more densely vegetated. A relatively small increase in the accumulation of Cecropia on the other hand, reveals minor expansion of forest in lowland humid habitats, probably into nearby disturbed sites (Fig. 6).

The vegetational record of zone P2 suggests a mild increase in rainfall as the Holocene began and somewhat moister conditions from ≈ 8,600 to 7,000 B.P (6.0 - 5.1 m). The following factors indicate that though moister than the late Pleistocene, the early Holocene climate remained drier than today: 1) persistence of abundant montane shrubs and xeric palms, 2) forest expansion limited to upland-moist and lowland-dry habitats, and 3) lack of development of lowland moist forests.

Hispaniola's record of dryness that persisted into the early Holocene is consistent with palynological evidence from the southeastern United States, where replacement of oak-dominated dry communities by moister vegetation commenced around 7,000 B.P. The dry climate however, is thought to have been warmer than at present (Watts, 1980; Watts and Stuiver, 1980). Fossil land snail faunas from Jamaica also indicate continued rainfall deficiency well into Holocene times (Goodfriend and Mitterer, 1988). In contrast, mainland sites in Central America and northern South America experienced rather moist conditions as the Holocene began (Leyden 1984, 1985; Piperno et al., 1990).

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Environmental Similarities of the Late Glacial
and Early Holocene

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Forest Fires and Vegetation

The sediments of zones P1 and P2 ($\geq$ 10,230 - 7,000 B.P.; 7.6 - 5.1 m) are characterized by low concentrations (and P2 also by low deposition rates) of charcoal fragments (Figs. 5, 8). The particles are small and worn, implying long distance transport. This evidence suggests occasional burning in the region throughout these episodes of severe drought. Natural burning of Hispaniola's modern vegetation occurs during the dry winter season (Holdridge, 1945), which is defined by a decrease in rainfall rather than temperature. Even highland, cool, moist forested sites are subjected to burning, though less often than drier lowland communities.

Under the modern climatic regime, average rainfall in arid sites of Hispaniola amounts to $\leq$ 850 mm/yr compared to 1,000 - 2,000 mm/yr in the lowlands of Haiti's southwestern peninsula (Woodring et al., 1924). These rainfall data together with the evidence of arid vegetation at Miragoane in the late Pleistocene, suggest that

mean annual precipitation at that time was about two thirds the modern average. Even during the early Holocene (P2) when precipitation began to increase, the environment was much drier than today. In both episodes the potential for natural burning was at least equal to that of present, and probably greater.

Data from later, wetter phases (zones P3 and P4) suggest a lack of climatic seasonality during the late Glacial and early Holocene. Thus, low incidence of fires is partly attributed to more even distribution of temperature and/or rainfall throughout the year. Climatic aseasonality could have reduced the rate of lightning strikes that spark spontaneous fires. Low litter accumulation, due in part to sparsity of herbs and the sclerophyllous, non-deciduous nature of prevalent xeric and shrubby plants, could have also contributed to reduce the incidence of natural fires. Studies of modern plant communities suggest that shrub environments may inhibit or slow fuel cover accumulation, thereby lowering fire frequency (Stephenson, 1990).

The pollen record of zones P1 and P2 indicates an expansion of montane elements, mostly shrubs but also <u>Podocarpus</u>, into areas lower than their modern range. Because montane taxa are fire-intolerant (Holdridge, 1947), the rarity of burning could have favored their downward displacement. This possibility, coupled with the likelihood that shrubs may be less moisture-demanding than their arboreal counterparts, could explain the abundance and ample distribution of the shrubby community throughout the late Pleistocene and early Holocene episodes. The environmental requirements of <u>Podocarpus</u> are not better known than those of the shrubs, though its occurrence in these episodes suggests resistance to drought.

Thermal Regime

The strong resemblance between the late Glacial and early Holocene pollen and charcoal records (P1 and P2) suggests climatic similarities, among them severe dryness and aseasonality. Rarity of fires during these two episodes was the main reason to suspect that the climate may have been also somewhat cooler than today. Because of the potential effects of other factors in suppressing burning, the infrequency of fires cannot be used as an indicator of the thermal conditions of these episodes and, unfortunately, the pollen record is uninformative in this respect. Nonetheless, some valuable clues are provided by the environmental history of the middle Holocene (zones P3 and P4).

The onset of Hispaniola's wettest conditions in the mid Holocene (beginning at 7,000 B.P.) was evidently associated with a warming trend. This contention is supported by evidence of rapid expansion of taxa such as <u>Cecropia</u>, typical of warm, humid habitats, when this period began (zones P2-P3 transition; Figs. 3, 6). Despite increased rainfall, lake levels began to decline slowly (Fig. 9) and fires became common (Figs. 5, 8). The occurrence of these two events at the inception of wet conditions in the mid Holocene, also points to the conclusion of climatic warming and suggests that the climate became more seasonal (e.g. with longer and/or more pronounced dry and wet seasons which may have differed in temperature, as they do at present). The climatic conditions inferred for the middle Holocene thus suggest that the preceding climate (in zones P1 and P2) was not only aseasonal and drier, but also cooler. However, it cannot be determined whether the mid-Holocene climate was as warm or warmer than at present.

Mean temperature differences of $\approx 4°C$ between the late Glacial and the Holocene have been estimated from studies of fossil Jamaican land snails (Goodfriend and Mitterer, 1988). It is not unreasonable to suspect that similarly

cool conditions affected Hispaniola when the Pleistocene ended. If, as postulated from Miragoane's record of forest growth, rainfall increases paralleled rises in temperatures, it can be concluded that in the earliest phase of the early Holocene (9,700 to 8,600 B.P.; 6.6 - 6.0 m) a small increase in both climatic factors occurred, whereas a more substantial increase took place between 8,600 and 7,000 B.P. (6.0 - 5.1 m). Lacking reliable fossil evidence of temperature change in the early Holocene record makes it difficult to ascertain if cool conditions in the Caribbean islands ended around 8,600 B.P. or if they lasted until 7,000 B.P. Both dates are within the time frame of termination of such conditions in neighboring continental lowland sites, assuming the chronologies of those places are correct.

Pollen data from the Panama Canal Zone suggest that two indicators of cool temperatures (Ericaceae and Symplocos) ceased to appear in the sediments around 8,500 B.P., whereas a third taxon, Iriartea (a stilt palm) persisted until ≈ 7,300 B.P., when modern warm temperatures were presumably attained (Bartlett and Barghoorn, 1973). Dated pollen assemblages from another Panamanian site (Piperno et al., 1990) and from Guatemala and Venezuela (Leyden, 1984, 1985), place the end of early Holocene cool conditions much earlier, at around 8,500 B.P. Discrepancies in the timing of termination of this cool event among these sites, may be a consequence of dating inaccuracies and/or to differences in the response of taxa to climatic change.

Relying on these chronologies, it is tentatively proposed that in Hispaniola too, cool temperatures ended around 8,600 B.P. (P2: 6.0 m level). Increasingly higher rainfall after this time, along with persistent aseasonality and low litter accumulation, probably account for the continued rarity of spontaneous burning up until 7,000 B.P. Another possibility is that the entire early Holocene zone (P2: 9,700 - 7,000 B.P.) records a cool period and that the current age of the zone's upper

boundary (5.1 m) is too young. Future research as well as redating of this core level might help to elucidate the timing of termination of the cool climatic period in the region.

It should be pointed out that at the 7.0 m level of the late Pleistocene-age zone (P1), the fossil record is rather suggestive of a brief warm, moist episode. At this level, there is a small but detectable increase in the percentage of Cecropia and Moraceae pollen (Fig. 3), a marked decrease in the proportion of Palmae and shrub pollen (Fig. 4), and high concentration of charcoal fragments (Fig. 5). A pronounced drop in $\delta^{18}O$ values (Fig. 9) suggests rising of the lake level, whereas development of a small littoral zone is indicated by a low percentage of Cheno/Ams. pollen, which otherwise is absent throughout the zone. At this core level, Moraceae is represented by pollen of Trophis-T, Chlorophora- T, and Pseudolmedia. The latter genus comprises 5% of the pollen sum in these deposits and is lacking in all other sediments below 4.2 m (except for a grain observed at 4.7 and at 5.3 m). Today, Pseudolmedia grows in mature, humid forests and along streams (T.A. Zanoni, personal comm.).

Charcoal concentration increases from 1,300/cm^3 at 6.9 m to 150,000/cm^3 at 7.0 m, while changes in pollen concentration are less pronounced: 78,000/cm^3 to 99,000/cm^3, respectively (Fig. 5). These data suggest that the increment in carbonized particles reflects higher incidence of burning, rather than heavy concentration of charcoal due to slow sedimentation.

The evidence of high lake level, slight expansion of tree taxa of warm, moist habitats (specially Pseudolmedia), and clear indications of increased burning, all point to the conclusion of a brief phase of temperature and moisture increase. In fact, some of the environmental indicators present at 7.0 m also typify the warm, wet mid Holocene episodes (zones P3 and P4). As the pollen record at 7.0 m does not

indicate profuse forest expansion, Pseudolmedia probably inhabited riparian forest. The time of occurrence and the duration of the brief moist, warm phase is unknown, but according to our current chronology it occurred during the late Glacial, prior to 9,700 B.P. (age at 6.7 m; see Table 2).

Middle Holocene (P3; and P4: 7,000 - 2,800 B.P.)

This Holocene interval is represented by two pollen zones: the Trema-Cecropia zone (P3) spanning from 7,000 to 4,800 B.P., and the Moraceae zone (P4) lasting from 4,800 to 2,800 B.P. Each zone is typified by pollen assemblages recording distinctive patterns of forest development. The following discussion summarizes characteristics common to both zones.

High charcoal concentrations document repeated burning of vegetation since fires became common around 7,000 B.P. (Fig. 5). The basal sediments of zone P3 (5.1 m), contain the earliest evidence of frequent burning of regional and local plant communities, provided by high deposition rates of carbonized plant fragments, mostly monocots, (Fig. 8) and abundant pollen of Ambrosia (Fig. 7), a genus not strictly indicative of burning but common in freshly disturbed, burnt sites (Tsukada and Deevey, 1967; Higuera-Diaz, 1983). Ambrosia pollen is nearly absent in zones P1 and P2, when burning was rare in the region.

Numerous Archaic-age remains have been found along the coast of Haiti's southwestern peninsula, but most of the artifacts are undated. Chronological sequences for the area place man's arrival in the peninsula as early as 6,000 B.P., or possibly later, until ≈ 4,000 B.P. Radiocarbon dating of samples from an archaeological site in Ile à Vache, an island off the peninsula's asouthern shore, yielded an age of 1,140 ± 50 B.C. (≈ 3,100 B.P.; Rouse and Moore, 1984). Hence, It is possible that human occupation of this part of Haiti postdates the inception of

high fire frequency documented in the fossil record of Lake Miragoane, beginning some 7,000 years ago.

At present, natural fires are common in the lowland drier and warmer parts of Hispaniola, but they also occur at higher elevation sites where precipitation is greater, though less frequently (Holdridge, 1947). Therefore, there is a good possibility that natural fires could also have occurred as the climate got warmer and seasonal in the mid Holocene, in spite of wetter conditions. I suspect that the inception of high incidence of burning at 7,000 B.P., was triggered by climatic change. For this reason, the time of initiation of this event should not be used to infer the time of arrival of humans to this region of Haiti, or anywhere else in Hispaniola. Until the time of the earliest occupation of southwestern Haiti becomes accurately known, human's role in initiating high incidence of burning in the region will remain questionable.

P3: Trema-Cecropia Assemblage Zone

Pollen zone P3 is encompassed by the 5.1 to 4.2 m section of the core (7,000 - 4,800 B.P.). Successional, mesic forests, often disturbed by fire, expanded around the deep lake, quickly replacing the palm and shrub-dominated communities. Expansion of upland and lowland moist forests is documented by high accumulation of Trema and Cecropia pollen, respectively (Figs. 6, 8). These genera are common second-growth trees at these sites (Holdridge, 1947). From this period on, Spondias and Bursera became regular components of the dry forest. This near-shore dry community also included other trees such as the fire-resistant Curatella and members of the Sapotaceae, and was rich in herbs, particularly Ambrosia and grasses (Fig. 7). Representation of other dry forest taxa like Myrtaceae and Guazuma has been sporadic from the beginning of this period until present.

The vegetational record of this episode strongly indicates a shift to a moister, warmer climate. In addition, the evidence of frequent fires (Fig. 5), presumably naturally-induced, and slowly declining lake levels (Fig. 9) in spite of high precipitation, suggest that the climate became more seasonal. The influence of newly established environmental conditions and competition with other taxa, apparently led to the eradication of the xeric, palm- dominated community from near the lake and to the displacement of montane shrubs from the mid-elevation sites where they once flourished.

P4: Moraceae Assemblage Zone

Sediments between 4.2 and 3.1 m (4,800 - 2,800 B.P.), comprise the Moraceae pollen zone. Hispaniola's most mesic Holocene vegetation is recorded in this episode by the rise in Moraceae pollen and by maximum accumulation of pollen from all other trees, except Trema (Figs. 6, 8). The dry, open forests typical of the lowlands surrounding the lake were dominated by Spondias, Curatella, and Ampelocera. Among the weeds, grasses and Ambrosia predominated (Fig. 7). The increased abundance of Spondias and Ampelocera might reflect proximity of these taxa to the basin. Each of these genera is about equally represented and when combined, their pollen acounts for half the total derived from all dry forest elements in this zone (Fig. 6). Conversely, other trees of this community such as Curatella, Bursera, the legumes and others whose rates of accumulation decline within the zone may have grown farther from the lake and/or become less common in the area.

The Moraceae forest, formerly represented by Trophis-T and Chlorophora-T, is also represented in this episode by pollen of Pseudolmedia, probably P. spuria, a species common along streams and in mature forests (T.A. Zanoni, personal

comm.). The three main genera of Moraceae are about equally represented while Ficus is as rare as in other zones of the core. The combined mean accumulations of Moraceae and Cecropia pollen ($\approx 2,000/cm^2/y$, each) account for more than half of the total pollen deposited in the sediments of this zone (Fig. 8).

Trema remains as the main highland forest representative but its abundance continues declining through zone P4 (Fig. 6). Alchornea, common in this period, and Zanthoxyllum represent the moist forest group. Pollen of Celtis, a deciduous secondary forest tree, begins to occur regularly in the sediments becoming very common in the upper part of zone P4 and in overlying zones. Celtis may be presently associated with the dry limestone mountain forests where deciduous taxa are prevalent (Howard, 1973), but due to its successional nature it may extend well beyond the realm of these forests, into drier lowland areas.

During these two millennia, mesic forests dominated in the lowlands and probably expanded into near-shore areas. Consequently, habitats for xeric species, whose extension was greatest in the late Pleistocene and early Holocene (zones P1 and P2), were severely reduced. Cecropia probably dominated in ecotones separating the dry and mesic forest of the warm lowlands, and in forest gaps at various elevations.

It is possible that during this episode of dense forests, the low elevation hills separating the Massif de La Selle from the Massif de La Hotte, may have been covered by mesic vegetation. The low-lying intervening area (≈ 150 km long) between the massifs currently supports dry forest. High accumulation of pine pollen (Fig. 6) reveals expansion of these forests, possibly encouraged by the high incidence of fires. Pinus occidentalis is presently more common on mountain summits. In the upland hardwood/pine forest ecotone, fire frequency largely controls the range of these communities. Where burning is rare, growth of fire- intolerant hardwood

species is favored. Pine populations expand into repeatedly burnt sites as this taxon is quite resistant to fire except when young (Holdridge, 1945, 1947). Fires in the highlands probably occur only every several years as opposed to more frequent burning in the drier lowlands. Frequent burning also favors expansion of pine into lowland areas but such populations are comparatively less dense than those of the summits.

<div align="center">Late Holocene (P5: 2,800 - 1,400 B.P.)</div>

This late Holocene episode presumably predates human disturbance of Miragoane's watershed and is represented by the <u>Celtis</u>-dry forest assemblage zone (P5), extending from 3.1 to 1.1 m. During this episode, the previously extensive mesic forests disappeared. Contemporary with this major reduction in forest cover there is evidence of a sharp drop in lake level and accelerated deposition of sediments in the basin. This set of environmental indicators suggests a shift to a much drier climate, apparently persisting at present.

The most prominent palynological change marking the beginning of this period is a 51% decrease in the average accumulation of arboreal pollen, from 5,700 to $2,800/cm^2$/yr. Moraceae's mean accumulation declined by 72%, whereas that of <u>Cecropia</u>, <u>Trema</u> and other mesic trees decreased by at least 40% (Fig. 8). <u>Celtis</u> reaches its maximum zonal mean accumulation in this episode and the dry forest attains its greatest taxonomic diversity. In addition to common, pre- existing taxa like <u>Curatella</u>, <u>Spondias</u>, <u>Bursera</u>, <u>Ampelocera</u>, legumes, etc., this forest is also represented by pollen of <u>Phyllostylon</u>, <u>Cordia</u>, <u>Sapindus</u>, some members of the Bombacaceae, Meliaceae (probably <u>Swietenia</u>), and several others. Maximum sedimentation rate, 0.11 g/cm^2/yr, for the entire core is recorded in this zone (Table 2, Fig. 5).

Simultaneous, pronounced fluctuations in pollen and charcoal concentrations (Fig. 5) suggest that the rate of sediment deposition was highly variable throughout this episode. Evidence of higher than average sedimentation is strongly suggested by very low microfossil concentrations in the bottom 70-cm and top 50-cm of the zone. The fossil-rich middle interval (2.4 -1.6 m) seems to represent a short phase of slower than average sedimentation. Minor, level-to-level fluctuations in fossil concentrations within each of these sections probably reflect short-term shifts in bulk sedimentation.

Pollen accumulation rates (Figs. 6, 7) were calculated assuming constant mean rate of sedimentation throughout the zone. Under this assumption, the diluting or concentrating effects of sedimentation rates higher or lower than the zone's mean, are not considered. Therefore, the large fluctuations in pollen accumulation rates among the various sections of zone P5 are partly an artifact of variable sedimentation.

Despite the effects of drastic changes in sedimentation on pollen accumulation rates, the pollen data suggest that the watershed became more forested during the time comprised by the 2.4 to 1.6 m section of zone P5. Except for Moraceae and Trema, the accumulation rates of pollen from most other tree taxa, and the weeds, show substancial increases in these deposits (Figs. 6, 7). Within the same core interval, the isotopic record indicates that the lake became temporarily deep (Fig. 9). Thus, both environmental indicators suggest that although the modern dry climate commenced by 2,800 B.P. (beginning of P5), there was a brief return to moist conditions sometime after 2,100 B.P (age at 2.2 m, Table 2). This moist climate lasted only a few centuries, and although apparently moister than the modern regime it was not nearly as mesic as the mid Holocene (zones P3 and P4).

The drastic fluctuations in lacustrine sedimentation within zone P5 are attributed to differerent factors. Fast sediment accumulation rate between 3.1 and 2.4 m, immediately after the massive retreat of nearby mesic forests, is undoubtedly related to intensified erosion of the newly exposed soils. The return of some forest to the watershed probably stabilized the local soils and sedimentation rates in the basin decreased during the time represented by the 2.4 to 1.6 m core interval.

CULTURE AND LOCAL VEGETATIONAL
DISTURBANCE AT MIRAGOANE

Taino Episode (P6: 1,100 - 500 B.P.)

Taino-age sediments correspond to the 72 to 30 cm section of the core. The beginning of this period coincides with the earliest known settlement of Miragoane's watershed by agriculturist people, the Tainos (Rouse and Moore, 1984; C. Moore, personal comm.), also customarily known as Arawak for the language they spoke (Rouse, 1948). Five Taino sites have been unearthed from the low hills about 1 km from the lake's southern shore. A few others lie at least 3 km away from the lake (C. Moore, personal comm.). Evidence of near-shore agriculture is provided by Zea pollen at 72 cm downcore (Fig. 4).

Plant associations characteristic of the preceding zone, P5, persist through this phase. Dry forest elements, pine, Cecropia, and Celtis are the dominant trees followed in importance by taxa from the relict moist forest (Figs. 6, 8). Pilea and Ambrosia either become rare or disappear in this zone, while grasses are much more abundant than before. Acalypha becomes common in these deposits and constitutes the bulk of the "other herbs" group (Fig. 7). Because grasses were already abundant in the previous zone, their increase in this episode cannot be attributed solely to human-induced disturbance of the local vegetation.

Disturbance of the local flora by the Tainos appears to have been minimal, at least when compared to the damage inflicted during Post-Columbian times. Taino agricultural acivities may have been confined to lightly forested areas or, as

previously presumed, the settlements may have been sparse and/or intermittent (Binford et al., 1987; Higuera- Gundy, 1989).

High deposition rates of pollen of aquatic plants (Fig. 7), mostly <u>Cladium</u> and <u>Typha</u>, document the early development and expansion of the extensive marsh connecting Lake Miragoane and its eastern neighboring lake, Petit Etang de Miragoane. Prior to ≈ 1,400 B.P. (1.1 m and below) these taxa were present but not abundant. The isotopic record of fast lake level decline in the past three millennia (Fig.11; see also Hodell et al., 1989, 1991 in review), suggests that the marsh area is a relictual shallow-water zone from past episodes of deeper water. In addition to falling lake water levels, the shallowing of this zone also may be related to accelerated inputs of land clastics since local deforestation began during Taino occupation of the watershed.

Sediments accumulated during the Taino episode are the most organic of all deep-water deposits so far recovered from the lake ($\overline{X}$ = 30%). The inorganic fraction contains equal, mean proportions of carbonates and silica (32%), and 5.1% iron oxide (Fig. 2, Table 1). Macrophytes and algal mats of the marsh probably have been an important source of organic matter to the lake since this community became established.

When sediment composition shifted from inorganic (in P5) to more organic in the Taino period (P6), the average dry density was reduced sixfold (to 0.14 g/cm^3) and mean dry mass accumulation rate dropped more than an order of magnitude, from 0.11 g/cm^2/yr to 0.01 g/cm^2/yr (Fig. 5). As a result, the accumulation rates of sediment components, expressed in terms of dry weight, also declined. In the Taino episode, organic matter, carbonate, and silica were deposited at a similar mean rate: 3 mg/cm^2/yr (Fig. 8), which is far slower than the values of deeper, high density deposits.

Post-Columbian Episode (P7: 500 B.P. - Present)

The uppermost 30 cm of Miragoane's cored sediments, represent this episode. The palynological and chemical changes characterizing this period have been interpreted as indicating massive deforestation and severe erosion of Miragoane's watershed during the past 500 years (Higuera- Gundy, 1989; Brenner and Binford, 1988; Brenner et al., 1988; Binford et al., 1987). Post-Columbian forests are mainly represented by successional taxa including Celtis, Cecropia, and the dry forest trees Sapindus and Bursera. As this period begins, the accumulation of arboreal and shrub pollen declines sharply (Figs. 6, 7) suggesting rapid initial clearance of large areas, probably for both agriculture and timber exploitation. Massive clearance of the local dry forest may be related to the fact that this type of community often provides conditions quite suitable for agriculture and livestock, and is easy to manage (e.g. slash and burn, etc.; Murphy and Lugo, 1986). Moreover, many of Miragoane's local dry forest trees are valuable timber species (Holdridge, 1945).

Erosion of the local soils is documented by a 50% increase in mean bulk sedimentation, to 0.015 $g/cm^2/yr$, and by changes in sediment composition (Figs. 5, 2). In Post-Columbian deposits, the proportion of organic matter is reduced to half ($\bar{X}$ = 15%) the values recorded in the Taino zone, and carbonate content increases twofold, to an average of 61% (Fig. 2, Table 1). Mean deposition of organic and siliceous materials is slightly slower in this zone than in P6, whereas carbonates accumulated three times faster (9 $mg/cm^2/yr$; Fig. 8) as deforestation accelerated in Post-Columbian times.

In the core interval from 8 to 6 cm, the concentration of pollen attains values similar to those recorded in Taino-age deposits (Fig. 5). In terms of accumulation rates, increases in Cecropia, Celtis, and the two main dry forest representatives

(Sapindus and Bursera) are the most prominent (Fig. 6). The sediment section encompassing this palynological change (10 to 6 cm), also shows an increase in organic matter relative to carbonates (Fig. 2). These data suggest temporary recovery of local forests, primarily to a nearly successional stage, and some degree of soil stabilization.

According to ^{210}Pb chronology, these sediments (8 to 6 cm) were deposited between 1852 and 1944 A.D. (M. Brenner, personal comm.). Even with these tentative dates, it is not possible to accurately determine the timing of forest recovery. It has been presumed however, that changes in land use following Haiti's independence from the French in 1804 may have favored this process (Higuera-Gundy, 1989; Binford et al., 1987). After independence, the large, French-run plantations were divided into smaller plots and used for subsistence agriculture. Small farms were created throughout the country and many highland sites were exploited for the first time (Woods, 1987; Holdridge, 1945). Return to subsistence agriculture and, perhaps, emigration of people from the watershed could have facilitated the brief recovery of forest documented by the fossil record.

The uppermost 6 cm of sediment record the past four decades of environmental history at Miragoane. Nearly devoid of pollen from all sources, these deposits document the most severe deforestation since the watershed was settled by agriculturists a thousand years ago. Intense erosion of the limestone terrain is reflected in higher carbonate content in the sediments and lower percentage of organic matter (Fig. 2).

At present, the vegetation in the drainage basin mainly consists of small patches of secondary forest trees separated by large expanses of land, some cultivated and some abandoned, sparsely covered by weeds. The bare soils are

severely eroded and those on the steep slopes of the south side of the lake, where the Tainos lived, have exposed bedrock.

It has been speculated that the severely deteriorated environment presently characterizing Miragoane's watershed, and most of Haiti, may resemble the environmental conditions faced by the Mayas in El Peten, Guatemala at the time of their cultural collapse around 900 A.D., after two and a half millennia of massive forest devastation (Brenner and Binford, 1988).

Riparian Disturbance and Lacustrine Sedimentation in the Miragoane Basin

Geochemical records documenting human-induced disturbance of lacustrine watersheds often, but not always, reveal increments in sediment accumulation rates as well as deposition of sediments lower in organic matter than those accumulated before near-shore human activity. This trend of change is well illustrated in the sedimentary record of several lakes in El Peten, Guatemala, where fast-paced accumulation of clay sediments, interbedded between more organic deposits, coincided with intense land use by the local Maya populations that inhabited the area from ≈ 3,000 to 400 B.P. (Deevey, 1978; Vaughan, et al., 1985; Brenner, in press).

Unlike the Guatemalan lakes, the record of Miragoane shows a tenfold decrease in mean bulk sedimentation and a fivefold enrichment in organic matter at the transition between pre-disturbance deposits (P5) and those accumulated during Taino times (P6; Figs. 2, 5). This unusual reversal in the geochemical record at the inception of human-induced disturbance, coincided with the develpment of the extensive marsh on the basin's eastern shore around 1,100 B.P.

There is an apparent cause-effect relationship between the establishment of macrophyte beds and the shift in sediment composition and rate of accumulation. It

is suspected that upon its establishment, the marsh may have intercepted substantial amounts of land-derived clastics, mostly inorganic, while increasing the inputs of organic matter to the lake. Consequently, as the loads of heavy inorganic materials declined and light-weight organic inputs increased, the sediment dry density and rate of deposition decreased. Even sediments deposited during the severe Post-Columbian disturbance (P7), accumulated at a much slower pace and are more organic than pre-disturbance deposits (zone P5 and below; see Figs. 2, 5). Compared with the Taino section (P6) however, Post-Columbian sediments accumulated 50% faster and are, on the average, only half as organic.

Evidence that marsh vascular plants, and quite likely their associated algal colonies, have been an important source of organic matter to the lake is provided by the sharp increase in organic content since the marsh community began to flourish at the site. The relatively high organic content (11 %) of recent sediments, 0 - 2 cm, which have been deposited during a time of severely eroded soils practically lacking an organic topsoil layer, also points to the marsh as the main source of organic material to the lake's deep-water zone. Insufficient time for diagenesis of these deposits can also account for their high organic content.

Geochemical analyses of littoral sediments from Lake Titicaca Menor, Bolivia, where macrophytes abound, suggest that these communities intercept and retain substantial amounts of suspended particles and dissolved nutrients when they pass from the land to the lake (Binford et al., in press). The chemical and palynological data from Lake Miragoane, record the development of macrophyte beds and associated sedimentary changes, and lend support to the claim that these communities can alter land-water transfer of materials. The data suggest that the littoral vegetation has not only intercepted incoming materials but has also contributed to organic enrichment of deep-water sediments. Through these

processes, the macrophyte stands have buffered the impact of accelerated soil erosion on the lake during anthropogenic disturbance.

PALEOECOLOGICAL HISTORY OF LAKE MIRAGOANE

The history of Lake Miragoane, based on $\delta^{18}O$ stratigraphy, is summarized in Figure 9. The oxygen isotopic data plotted in the Figure are according to Hodell et al. (1989, 1991 in review). In Lake Miragoane, a nearly closed lacustrine system, the variations in $\delta^{18}O$, observed in fossil ostracod shells, are interpreted as recording lake level fluctuations induced, chiefly, by changes in evaporation/precipitation ratios (Hodell et al., 1989, 1991 in review). Evaporation naturally favors the loss of the lighter and more abundant ^{16}O isotope, thus causing ^{18}O-enrichment of the remaining water (Hodell et al., 1991; Covich and Stuiver, 1974).

Late Pleistocene (P1: > 10,230 - 9,700 B.P.)

Peak $\delta^{18}O$ values recorded during the late Pleistocene (P1: 7.6 - 6.6 m), suggest low lake levels and high evaporation/precipitation (E/P) ratios during this dry episode (Hodell et al., 1989, 1991 in review). Evidence that the shallow lake was a permanent fresh-water body is provided by a continuous record of lacustrine sediments and by presence of <u>Candona</u> sp., a fresh-water ostracod found at all levels in the core.

Palynological and chemical data from a 1.1 m core raised in 1983 from the lake's southern bay (Fig. 1), shed some light regarding Lake Miragoane's water level when the Pleistocene ended. In that core, the sediments below 80 cm are organically rich (up to 30%), contain abundant pollen of shrubs (up to 90%) and lack pollen of Cheno/Ams. (for plots showing these data see Binford and Deevey, 1985). At the 80 cm level, pollen of Cheno/Ams. suddenly appears in large

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Figure 9. $\delta^{18}O$ stratigraphy of Lake Miragoenae sediments (modified from Hodell et al., 1991, in review). The record of lake water level fluctuations is based on these isotopic data.

LAKE MIRAGOANE, HAITI

δ 18O (‰, PDB)
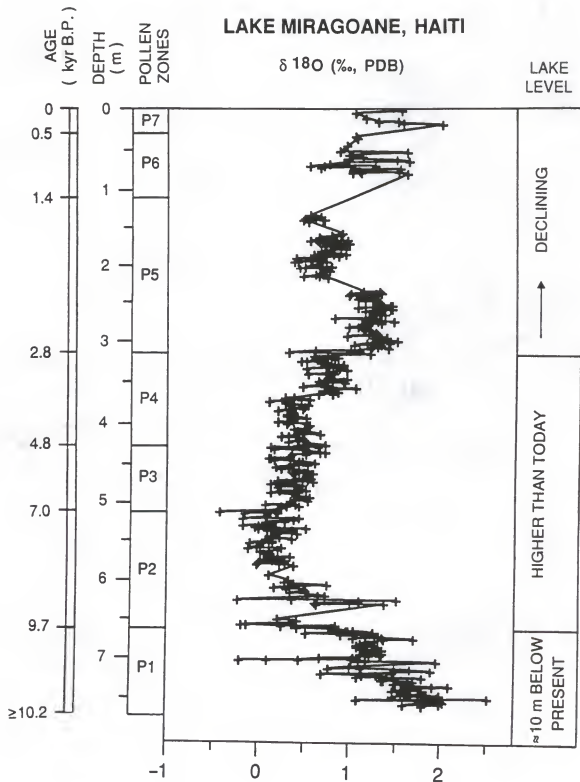
numbers (up to 30%), shrub pollen representation declines (to ≤ 40%), and organic matter decreases sharply (to ≤ 5% through the rest of the core). The Cheno/Ams.-shrub pollen stratigraphic shift at 80 cm, resembles the change characterizing the late Pleistocene-early Holocene transition (P1/ P2) in the core studied in this work (raised in 1985). The nature of the palynological change recorded in the two cores suggests that the shift was synchronous and occurred around 9,700 B.P., according to the chronostratigraphy of the 1985 core (see Figs. 4 and 7).

Based on this stratigraphic relationship, the organic-rich 30 cm of sediment at the bottom of the 1.1-m core were deposited during the late Pleistocene, while those above the 80 cm level are of Holocene age. The bay site where the core was raised is currently under 10 m of water. The high organic content of the late Glacial-age muds from the bay suggests that at that time this site was a highly productive shallow-water zone. Support to this claim is provided by the low organic content of deep-water sediments (1985 core) deposited during the same time interval (see zone P1 in Fig. 2). Based on the chemical evidence of the 1.1 m core, it is estimated that the lake level was ≈ 10 m lower than today when the Pleistocene ended (9,700 B.P.), but may have been lower before then.

A sea level curve for the western Atlantic and Caribbean region suggests that in the late Glacial the level of the sea was ≈ 30 m below present (Taylor et al., 1985). Hence, Miragoane's water surface was ≈ 40 m a.m.s.l. Consequently, the basin was not a cryptodepression at that time (i.e. little or no portion of it was below sea level as has been the case since the Holocene began).

Early Holocene (P2: 9,700 - 7,000 B.P.)

Minimum isotopic values were attained and sustained through much of the early Holocene indicating water levels higher than those of the modern lake (Hodell et al., 1989, 1991 in review). The sharp drop in $\delta^{18}O$ at the beginning of this phase suggests rapid filling of the lake. Though higher than at present, the lake level briefly dropped between ≈ 9,200 and 8,750 B.P. (6.35 - 6.1 m) and then rose again (Fig. 9).

Filling of the Miragoane basin coincided with the development of an extensive littoral zone dominated by weedy species of Cheno/Ams. (Figs. 4, 7). Shallow-water habitats for these plants persisted throughout this episode and may have been created by gradual rising of the water over gentle basin slopes or, as the isotopic record seems to indicate, by rapid flooding of large, flat, marginal areas.

Initial deepening of lake Miragoane is partly attributed to the increase in rainfall indicated by the palynological record. Nevertheless, because the basin is located in lowland karst terrain near the ocean it is likely that the rapid early Holocene sea level rise (Taylor et al., 1985; Mann et al., 1984) raised the water table, and the lake level. Rising water levels of low-lying lakes located in similar terrain in neighboring Florida, U.S.A. and Yucatan, Mexico, are thought to have been strongly influenced by the early Holocene eustatic sea level rise (Watts, 1980; Covich and Stuiver, 1974).

The fact that soon after the lake filled, water levels quickly declined, between 9,200 and 8,750 B.P. (6.35 to 6.1 m), agrees with the interpretation of strong marine control on the initial filling of the basin. The pollen record suggests continued rainfall deficiency up until ≈ 8,600 B.P., which in combination with high evaporation/precipitation ratios may account for the temporary subsidence of the lake level. Return to very deep conditions after 8,750 B.P. appears related to

increased rainfall, indicated by pollen evidence of progressive forestation of the watershed.

Sediment accumulation rate amounted to 0.07 g/cm^2/yr from 9,700 to 9,000 B.P. (6.7 - 6.2 m) and then decreased to 0.04 g/cm^2/yr until 7,000 B.P. (Table 2). The average sedimentation rate for the zone equals 0.05 g/cm^2/yr (Fig. 5). The early phase of fast sedimentation most likely resulted from heavy erosion of nearby soils, as near-shore areas were only sparsely forested. Slumping and redeposition of materials from marginal sites into deep-water zones probably occurred as the lake filled. Despite presence of a large littoral community, organic sedimentation is very low (2.8 mg/cm^2/yr; Fig. 8). It seems as if most of the organic matter derived from marginal plants was quickly decomposed and/or deposited in situ, rather than transferred to deep- water sites.

When the lake level rose at the beginning of the Holocene, Miragoane and Petit Etang de Miragoane, a smaller basin located $\approx$ 2.5 - 3 km to the east, probably became hydrologically connected to form a single, large lake. The connecting area, currently occupied by a marsh, may have been a few meters deeper (in terms of higher water level and lower floor) throughout the deep-water phases (zones P2 through P4). Isotopic evidence of rapidly falling lake levels in the past 3,700 years also implies progressive desiccation of the hydrological connection, and indirectly suggests that the area eventually became shallow enough to support a marsh community since $\approx$ 1,100 B.P.

The isotopic data were used to determine historical lake level changes, from shallow in P1 to very deep in P2. Before δ^{18}O results and radiocarbon dates were available, the absence of Cheno/Ams. in P1 was erroneously interpreted as indicating a deep-water stage, and their abundance in P2 as revealing subsiding lake levels (Higuera-Gundy, 1988). The inference was supported by the near lack of evidence of fires in both zones, whose rarity was assumed to be related to a moist

early Holocene climate (believed to be represented by zone P1, now confirmed to represent late Glacial deposits).

Middle Holocene (P3 and P4: 7,000 - 2,800 B.P.)

Isotopic data indicate progressive shrinking of the lake throughout the two middle Holocene wettest episodes (P3 and P4; 5.1 - 3.1 m), even thoguh levels higher than today were maintained (Hodell et al., 1989, 1991 in review). Climatic warming and the onset of seasonality beginning at 7,000 B.P., probably account for the drop in lake level, as high evaporation/precipitation ratios ensued during the drier months. When water levels gradually declined, through zones P3 andP4, habitats for littoral plants were reduced, as suggested by marked drops in the accumulation of Cheno/Ams. (Figs. 7, 8). Forestation of the watershed evidently stabilized the local soils and as a result, allogenic inputs and lacustrine sedimentation subsided (Figs. 5, 8).

Late Holocene (P5, P6, P7: 2,800 B.P. - Present)

Based on our current chronology and the $\delta^{18}O$ record (Fig. 9), modern desiccation of the lake began around 3,700 B.P. (3.6 m level of zone P4), but water losses have been more pronounced during the past 2,800 years of climatic dryness. Despite evidence of a prevalently dry climate during the last three millennia, both the isotopic and palynological records suggest a return to moister conditions during the time interval encompassed by the 2.4 to 1.6 m section of the core. Within this interval of zone P5, the $\delta^{18}O$ values drop (Fig. 9) suggestig rising of the lake level, while pollen data (Figs. 6, 7) document certain degree of forestation in the watershed.

Above 1.5 m, the isotopic record indicates shrinking but fluctuating lake levels, and in some instances suggests temporary rises in water level. In general, the

isotopic data reveal that Lake Miragoane, like other lowland tropical lakes such as Enriquillo in the Dominican Republic, Valencia in Venezuela, and Laguna de Chichancanab in Yucatan, Mexico, among others, has been experiencing substantial desiccation within the past three thousand years (Taylor et al., 1985; Mann et al., 1984; Bradbury et al., 1981; Covich and Stuiver, 1974; Lewis and Weibezahn, 1981).

Low pollen and charcoal concentrations in the topmost 50 cm of zone P5 (1.6 - 1.1 m; Fig. 5), led to conclude fast sediment deposition rates within this core interval. Fast lacustrine sedimentation could have resulted from accelerated delivery of materials derived from the marsh area, prior to the establishment of macrophyte beds. At the time these sediments were deposited, the lake level was rapidly declining. Up until about 1,400 B.P. (end of zone P5), the water level of the low-lying marsh area might still have been sufficiently deep to prevent profuse growth of macrophytes but shallow enough for its bottom sediments to undergo strong mixing by wind and currents, thus promoting the transfer of large amounts of sediments from this source into deeper parts of the basin.

WEST INDIAN BIOGEOGRAPHY AND LATE QUATERNARY
ENVIRONMENTS: A GENERAL APPROACH

Modern biogeographic patterns in the Caribbean islands undoubtedly reflect the impact of repeated environmental and geological changes predating the late Pleistocene. However, the climatic and vegetational changes indicated by the pollen record of Hispaniola appear to have been sufficiently intense to cause further modification of the insular biota within the past ten thousand years. Because animals strongly depend on vegetation for food, habitat, and shelter, it is suspected that shrinking and expansion of various plant communities could have severely diminished the availability of habitat and food resources for their associated faunas and, in some instances, could have led to the demise of some species.

Morgan and Woods (1986) have outlined three episodes of major faunal extinction in the Antilles: late Pleistocene-early Holocene (40,000 - 4,500 B.P.), Amerindian (4,500 - 500 B.P.), and Post-Columbian (500 B.P.- Present). They acknowledge the lack of radiocarbon dates for most of the West Indian fossils, so that the timing of species extinction is a rough approximation. The lost species include numerous land mammals, birds, reptiles, as well as bats and other Antillean vertebrates (Morgan and Woods, 1986; Morgan, 1989; Woods, 1989a, 1989b; Pregill and Olson, 1981; Steadman et al., 1984). Miragoane's chronology of major changes in climate and vegetation roughly coincides with the proposed time of extinction of faunal taxa. Such contemporaneity suggests, but does not prove, a cause-effect relationship between the two events.

Late Glacial and early Holocene decimation of West Indian vertebrates and reduced ranges of those that survived, are generally attributed to climatically-induced alteration of habitats (from arid to more mesic), and to habitat elimination and/or fragmentation as the sea level rose (Pregill and Olson, 1981). One of the most interesting examples of past, severe alteration of dry habitats is given by the distribution of several obligate xerophilic species, specially reptiles and birds. According to Pregill and Olson (1981), living members of these groups have disjunct or restricted distributions and oftentimes their fossils occur in areas outside their present range, thus indicating a much wider distribution during times when xeric vegetation was more widespread, such as in the late Glacial.

The palynological evidence is conclusive regarding the existence of larger-than-present distribution of xerophytic communities during the late Pleistocene as well as the early Holocene (zones P1 and P2; $\geq$ 10,230 - 7,000 B.P.). It also suggests severely reduced and fragmented moist forested habitats, thereby implying wide disjunctions among mesophilic groups. Small islands and those with gentle low-lying topography, were probably dominated by xeric plant communities and had smaller areas with moist forest trees than Hispaniola, at least during the late Pleistocene.

Contraction and fragmentation of xerophytic communities and, quite likely, their associated faunas, inevitably occurred as mesic forests grew and expanded throughout the island in the mid Holocene (P3 and P4; 7,000 - 2,800 B.P). Shrinking of Hispaniola's dry habitats some 7,000 years ago, coincided with the onset of high incidence of fires, presumably induced by natural factors. The potential impact of fires on the Antillean fauna should be considered in future zoogeographic studies. The inception of high frequency of forest fires in other Caribbean islands however, may or may not have been contemporaneous with Hispaniola's.

Mid Holocene expansion and coalescence of mesic forests probably favored contact between previously isolated mesophilic groups of plants and animals. It is within the realm of possibility that a transect of lowland dry forest ($\approx$ 150 km in length) currently separating the Massif de La Hotte from the Massif de La Selle, could have been occupied by mesic forests (successional and/or mature) for at least part of the time when such communities were prevalent between 7,000 and 2,800 B.P.

In this locality, three coexisting species of the mammalian insectivore Nesophontes (that became extinct $\approx$ 100 years ago; C.A. Woods, personal comm.), show various degrees of morphological separation in the massifs and their intervening area. The morphological differences suggest recent contact in the Miragoane region between formerly isolated populations from the massifs (information provided by C.A. Woods to M.W. Binford; see Binford and Deevey, 1985). The time of contact between the two populations is unknown, but if it happened in the Holocene, the most favorable period was during the time when mesic vegetation presumably closed the wide gap formerly separating the massifs.

Elimination of some vertebrate species during the Amerindian period (4,500 B.P. - Present), is suspected to have resulted from direct exploitation and destruction of habitats by man, and more recently also by the introduction of exotic species such as Rattus (Morgan and Woods, 1986), and domestic cats and dogs (C.A. Woods, personal comm.). Although humans arrived in the Greater Antilles around 7,000 B.P. (Cruxent and Rouse, 1969; Rouse, 1989), their role as the leading cause of animal extinctions in the past 4,500 years may be overestimated.

Data from southwestern Hispaniola indicate that several mammal species (including rodents, ground sloths, and a primate) went extinct around 3,000 B.P. (Woods, 1989a, 1989b). Fossils of species documenting these extinctions were

recovered from sinkholes located in wet, forested highland sites in the Massif de La Hotte, and other sites in the Massif de La Selle (Woods, 1989a). A faunal assemblage taxonomically similar to the one found in the sinkholes was unearthed from a cave located at 250 m elevation in the Massif de La Hotte. Radiocarbon dates from fossils in the cave are 10,320 and 3,715 B.P. However, it is suspected that the cave fossils may have been accumulated by a large barn owl (Tyto ostologa, now extinct) whose remains also occurred in the cave (Woods, 1989a). Hence, the immediate vicinities of the lowland cave may not have been the natural habitat for the mammal assemblage under consideration. In general, the non-volant mammalian fossil fauna of the massifs suggests that of the endemic rodents known to Hispaniola, 93% are now extinct but most survived until ≈ 3,000 B.P. (Woods, 1989a, 1989b). Within the past few millennia, four species of bats have also gone extinct from the highlands of the Massif de La Selle (Morgan and Woods, 1986).

It is still uncertain whether humans arrived in Haiti's southwestern peninsula as early as 6,000 B.P. or much later, about 4,000 B.P. (Rouse and Moore, 1984). In either case, as these early settlers strongly depended on the sea for food, they mainly inhabited coastal sites (Rouse, 1948). Their disturbance of the local flora and fauna probably did not reach far into the highlands of the massifs. Agriculturists, who also made pottery, came to the peninsula about 1,400 B.P., and their archaeological remains, too, have been recovered from the lowlands (Rouse and Moore, 1984). Based on this archaeological information and the fact that extinction of some highland mesophilic animal species was coincident with severe loss of mesic forests (by 2,800 B.P.), it is not unreasonable to presume that deterioration of moist habitats played an important role in the demise of these taxa.

The impact of humans on plant and animal communities must have varied, in time and intensity, from island to island. For this reason, it is pertinent to use the
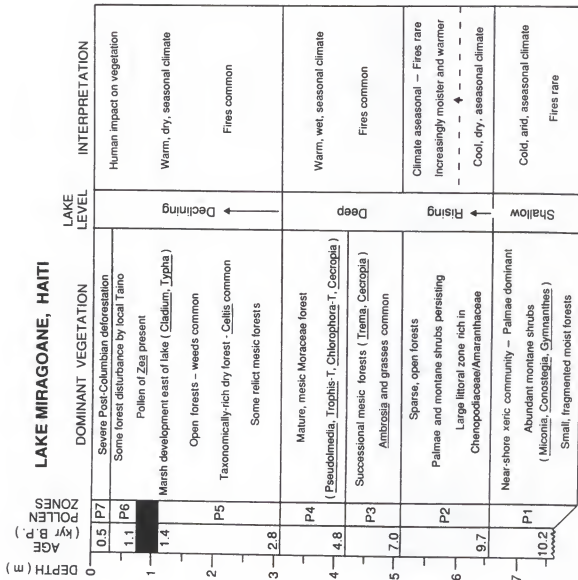
available environmental records in conjunction with archaeological data for individual islands, in order to better understand the role of man and/or natural events in altering the islands' flora and fauna.

SUMMARY AND CONCLUSIONS

The environmental history of Lake Miragoane is summarized in Figure 10. The fossil record agrees with findings from others sites indicating widespread aridity in the lowland tropics during the late Pleistocene. In Hispaniola, the late Glacial and early Holocene (P1 and P2; $\geq$ 10,230 - 7,000 B.P.) climate was aseasonal, and drier and cooler than at present., though cool temperatures appear to have ceased by 8,600 B.P. Deficient rainfall and edaphic aridity (brought about by rapid water loss through the limestone terrain) during these two periods, favored a large expansion of dry communities over the island, with xeric vegetation dominating in the lowlands and shrub species (e.g. of Melastomataceae and Euphorbiaceae) in mid- and high-elevation slopes, where moisture was probably greater. On mountain sites near the basin, small populations of moist forest trees existed either mixed with the shrubs and/or restricted to climatically favorable habitats. In general, the island's moist forests were limited in area and, quite likely, fragmented.

Climatic and non-climatic factors contributed to suppress natural fires. Low temperature and the infrequency of burning appear to have favored the descent of fire-intolerant montane elements, mainly shrubs, into areas lower than their modern range. Because of aridity in the lowlands however, the downward displacement of montane taxa may have been less than expected from the modern lapse rate of temperature change with elevation (1°C/150 m; Woodring et al., 1924). In rainshadow areas, such as southern Haiti and more generally, on leeward mountain slopes, xerophytic communities probably expanded far above their current limits. Islands with smaller area and/or less pronounced topographic relief than

Figure 10. Summary of the paleoecological history of Lake Miragoane, Haiti, in relation to changing climate and the impact of local riparian disturbance by man. Blackened area (0.72 - 1.1 m) indicates core hiatus.

LAKE MIRAGOANE, HAITI

DEPTH (m)	AGE (kyr B.P.)	POLLEN ZONES	DOMINANT VEGETATION	LAKE LEVEL	INTERPRETATION
	0.5	P7	Severe Post-Columbian deforestation	Declining	Human impact on vegetation
	1.1	P6	Some forest disturbance by local Taíno		
	1.4		Pollen of Zea present		Warm, dry, seasonal climate
1			Marsh development east of lake (Cladium, Typha)		
2		P5	Open forests – weeds common		Fires common
			Taxonomically-rich dry forest – Celtis common		
3	2.8		Some relict mesic forests		
		P4	Mature, mesic Moraceae forest (Pseudolmedia, Trophis-T, Chlorophora-T, Cecropia)	Deep	Warm, wet, seasonal climate
4	4.8				Fires common
5		P3	Successional mesic forests (Trema, Cecropia) Ambrosia and grasses common		
	7.0			Rising	Climate aseasonal – Fires rare
6		P2	Sparse, open forests Palmae and montane shrubs persisting Large littoral zone rich in Chenopodiaceae/Amaranthaceae		Increasingly moister and warmer Cool, dry, aseasonal climate
7	9.7		Near-shore xeric community – Palmae dominant	Shallow	
	10.2	P1	Abundant montane shrubs (Miconia, Conostegia, Gyrinanthes) Small, fragmented moist forests		Cold, arid, aseasonal climate Fires rare

Hispaniola, may have primarily supported xeric vegetation and very little forest when the Pleistocene ended.

Both temperature and rainfall increased as the Holocene began (9,700 B.P.), but the climate remained comparatively drier than at present until 7,000 B.P. Early Holocene-age evidence from various sites indicates that the amount of rainfall over the Caribbean region differed greatly. While dry conditions persisted in Hispaniola, Jamaica (Goodfriend and Mitterer, 1988), and the southeastern U.S.A. (Watts, 1980; Watts and Stuiver, 1980), the climate of Central America and northern South Ameica was rather mesic (Leyden, 1984, 1985). Assuming accurate dating of the fossils from these sites, their evidence raises the possibility that the Greater Antilles, and perhaps some of the northern Lesser Antilles, could have been much drier than other islands farther south, such as the Lesser Antilles closer to the mainland. This conjecture however, cannot be corroborated by Miragoane's record.

A high-resolution $\delta^{18}O$ record from Lake Miragoane, documents short and long-term fluctuations in water level related to marked variations in evaporation/precipitation ratios (Hodell et al., 1989, 1991 in review). During the late Glacial, Miragoane was a fresh-water lake, but its water level was 10 to 11 m lower than today by the end of that period (9,700 B.P.). Lake Miragoane and Lake Enriquillo, of similar depth but saline, may have been the only basins in Hispaniola with permanent water, as most others in the island are presently rather shallow (Bond, 1935). At the onset of the Holocene (9,700 B.P.) the Miragoane basin quickly filled, influenced by higher rainfall and rapid sea level rise. Water levels much higher than today were maintained until ≈ 2,800 B.P., when a new episode of desiccation began (Fig. 9).

Hispaniola's most mesic vegetation flourished around the deep lake during the mid Holocene (zones P3 and P4; 7,000 - 2,800 B.P.), when the climate turned warm,

wet, and seasonal. Since the inception of this climatic regime, forest fires have been common in the island. Development of mesic semi- evergreen forests proceeded in two steps. The early stage of forest growth (P3; 7,000 - 4,800 B.P.) commenced with the rapid proliferation and expansion of fire-disturbed secondary communities, dominated by Cecropia in the lowlands and by Trema in the uplands. The forestation process may have substantially improved the soils' quality.

Optimal conditions for the development of mature forest prevailed between 4,800 and 2,800 B.P. (P4), Hispaniola's most mesic episode. Within this time period, closed forests dominated by Moraceae trees occupied vast lowland areas, thereby severely limiting the distribution (and possibly leading to fragmentation) of dry forest and other dry communities typical of these sites. Although coalescence of mesic communities in the mid Holocene probably favored contact between formerly disjunct groups of mesophilic plants and animals, the simultaneous introduction of frequent fires could have imposed some limitations on the distribution and/or survival of some species.

Return to dry conditions in the late Holocene (2,800 B.P.- Present) led to the decimation of the ancient mesic forests and gave way to modern plant communities, typified by semi-deciduous and relictual semi-evergreen forest taxa and abundant weeds. The dry climate has contributed to rapid desiccation of the lake (Hodell et al., 1989, 1991 in review). Although basically a dry climate has prevailed in the past 3,000 years, conditions moister than at present but less mesic than in the mid Holocene existed for a few centuries after ≈ 2,200 B.P. During this brief moist episode (recorded between 2.4 and 1.6 m of the core), the lake level rose and near-shore populations of some dry forest trees and Cecropia increased.

A hydrological connection between Miragoane and Petit Etang de Miragoane, formed around 9,700 B.P., is now a relictual shallow-water zone occupied by a

marsh established about the time of man's arrival in the watershed ≈ 1,100 B.P. or slightly earlier. This community has probably functioned as a retention area of heavy incoming clastics and as a source of organic matter to deep parts of the lake. Hence, the sediments deposited during human occupation are organic-rich and accumulated at slower rates than those in pre- disturbance core sections (1.1 m and below).

During Taino occupation of Miragoane's watershed (1,100 - 500 B.P), very little vegetation disturbance seems to have occurred. The environmental history of Haiti in Post-Columbian times (500 B.P. - Present) probably parallels that of Miragoane, in that massive deforestation and consequent soil erosion have been nearly continuous since the Spanish conquest. A brief episode of recovery of Miragoane's forest and soils (recorded between 8 - 6 cm), has been attributed to changes in land use and, possibly, emigration of people from the lake's watershed soon after Haiti's independence from the French in 1804. In recent decades, widespread devastation of Haiti's natural resources has been intensified by the pressure of large population densities.

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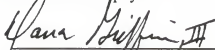
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BIOGRAPHICAL SKETCH

Antonia Higuera-Gundy is a Mexican citizen, born in Ignacio Zaragoza, Tamaulipas, on January 20, 1948. She obtained her Biological Sciences Bachelor degree from the Universidad de Nuevo Leon (located in Monterrey, N.L., Mex.) in June, 1974. As a partial requirement for this degree, she wrote a thesis entitled, "Polenes anemofilos mas comunes en el area metropolitana de Monterrey, N.L., Mex.," under the supervision of Dr. Carlos Canseco Gonzalez.

From 1971 to 1978, she worked at the Universidad de Nuevo Leon, teaching undergraduate courses in botany, biology, and other related subjects. Mrs. Higuera-Gundy enrolled in the Botany Department, University of Florida, to pursue her graduate career. Under the supervision of Dr. E.S. Deevey, she began her studies on tropical paleolimnology. Her Master of Science thesis, entitled "A paleolimnological record of human disturbance in lakes Atitlan and Ayarza, Guatemala," was presented in 1983. She has published or coauthored several papers from the data presented in this dissertation and from her master's thesis. Other related activities include seminar presentations and some consulting work in palynology.
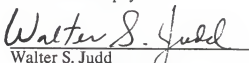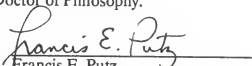
I certify that I have read this study and that in my opinion it conforms to acceptable standards of scholarly presentation and is fully adequate, in scope and quality, as a dissertation for the degree of Doctor of Philosophy.

Dana Griffin III, Chair
Professor of Botany

I certify that I have read this study and that in my opinion it conforms to acceptable standards of scholarly presentation and is fully adequate, in scope and quality, as a dissertation for the degree of Doctor of Philosophy.

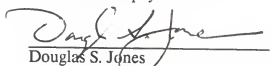Frank G. Nordlie
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This dissertation was submitted to the Graduate Faculty of the Department of Botany in the College of Liberal Arts and Sciences and to the Graduate School and was accepted as partial fulfillment of the requirements for the degree of Doctor of Philosophy.

August 1991

Dean, Graduate School